Geochronology and stratigraphy of late Pleistocene lake cycles on the southern Bolivian Altiplano: Implications for causes of tropical climate change

Christa Placzek†
Jay Quade
P. Jonathan Patchett
Department of Geosciences, University of Arizona, Tucson, Arizona 85721, USA

ABSTRACT

Large paleolakes (~33,000–60,000 km^2) that once occupied the high-altitude Poopo, Coipasa, and Uyuni Basins in southern Bolivia (18–22°S) provide evidence of major changes in low-latitude moisture. In these now-dry or oligosaline basins, extensive natural exposure reveals evidence for two deep-lake and several minor-lake cycles over the past 120 k.y. Fifty-three new U-Th and 87 new ^{14}C dates provide a chronologic framework for changes in lake level. Deposits from the "Ouki" deep-lake cycle are extensively exposed in the Poopo Basin, but no deep lakes are apparent in the record between 98 and 18.1 ka. The Ouki lake cycle was ~80 m deep, and nineteen U-Th dates place this deep-lake cycle between 120 and 98 ka. Shallow lakes were present in the terminal Uyuni Basin between 95 and 80 ka (Salinas lake cycle), at ca. 46 (Inca Huasi lake cycle), and between 24 and 20.5 ka (Sajsi lake cycle). The Tauca deep-lake cycle occurred between 18.1 and 14.1 ka, resulting in the deepest (~140 m) and largest lake in the basin over the past 120 ka. Multiple ^{14}C and U-Th dates constrain the highest stand of Lake Tauca along a topographically conspicuous shoreline between 16.4 and 14.1 ka. A probable post-Tauca lake cycle (the Coipasa) produced a ≤55-m-deep lake that is tentatively dated between 13 and 11 ka.

We suggest that paleolakes on the Bolivian Altiplano expanded in response to increased moisture in the Amazon and enhanced transport of that moisture onto the Altiplano by strengthened trade winds or southward displacement of the Intertropical Convergence Zone (ITCZ). Pole-to-equator sea-surface temperature (SST) and atmospheric gradients may have influenced the position of the ITCZ, affecting moisture balance over the Altiplano and at other locations in the Amazon Basin. Links between the position of the ITCZ and the ca. 23 ka precessional solar cycle have been postulated. March insolation over the Altiplano is a relatively good fit to our lake record, but no single season or latitude of solar cycling has yet to emerge as the primary driver of climate over the entire Amazon Basin. Temperature may influence Altiplano lake levels indirectly, as potentially dry glacial periods in the Amazon Basin are linked to dry conditions on the Altiplano. Intensification of the trade winds associated with La Niña–like conditions currently brings increased precipitation on the Altiplano, and deep-lake development during the Tauca lake cycle coincided with apparently intense and persistent La Niña–like conditions in the central Pacific. This suggests that SST gradients in the Pacific are also a major influence on deep-lake development on the Altiplano.

Keywords: lakes, U-Th, climate, Bolivia, insolation, ENSO.

INTRODUCTION

The tropics are implicated in forcing global climate shifts at interannual to orbital time scales (e.g., Cane, 1998), and the global impact of El Niño–Southern Oscillation (ENSO) variability is one example of how the tropical moisture can propagate rapid changes worldwide. The scarcity of well-dated paleoclimate records from low latitudes, however, leaves open many questions about the interplay between tropical and global climate. Many now recognize that glacial cooling in the tropics was as much as 5–8 °C (e.g., Stute et al., 1995; Thompson et al., 1995; Colinvaux et al., 2000), but during global glaciations, the magnitude of hydrologic changes in the Amazon Basin, an ecosystem with large potential impacts on global biogeochemical cycles, remains disputed. A number of biologic proxies suggest little or no change in the tropical forest during the late Pleistocene (Haberle and Maslin, 1999; Colinvaux et al., 2000; Kastner and Goñi, 2003). By contrast, pollen records (Behling, 2002), clay mineralogy of Amazonian sediment (Harris and Mix, 1999), and some models (e.g., Garreaud et al., 2003) suggest relative aridity over this same interval.

A long-range lake record from the southern Bolivian Altiplano can serve as a critical test of the relationship between tropical moisture and global climate. The Bolivian Altiplano (14–22°S) is a large, high-elevation (~3800 m) plateau situated at the western end of the Amazon rainfall belt. The South American summer monsoon, which brings copious rains to the Amazon Basin, is responsible for more than 80% of the annual rainfall on the Altiplano (Vuille, 1999). Over long time scales, the position of the Altiplano at the distal end of the Amazon rainfall belt makes this location sensitive to relatively modest changes in climate. Modern interannual variability on the northern Altiplano is also strongly tied to ENSO, which modulates the strength of the trade winds, generally causing El Niño years to be dry (Vuille, 1999).

Most existing paleoclimate studies (e.g., Baker et al., 2001a, 2001b; Seltzer et al., 2000; Rowe et al., 2002; Fritz et al., 2004) point to local insolation changes as the cause of rainfall variability over the Altiplano. Altiplano lake-level records, based largely on proxy evidence from cores, provide crucial support for the idea that the precessional solar cycle is the principal

†E-mail: cplaczek@geo.arizona.edu.

GSA Bulletin; May/June 2006; v. 118; no. 5/6; p. 515–532; doi: 10.1130/B25775.1; 13 figures; 3 tables; Data Repository item 2006100.

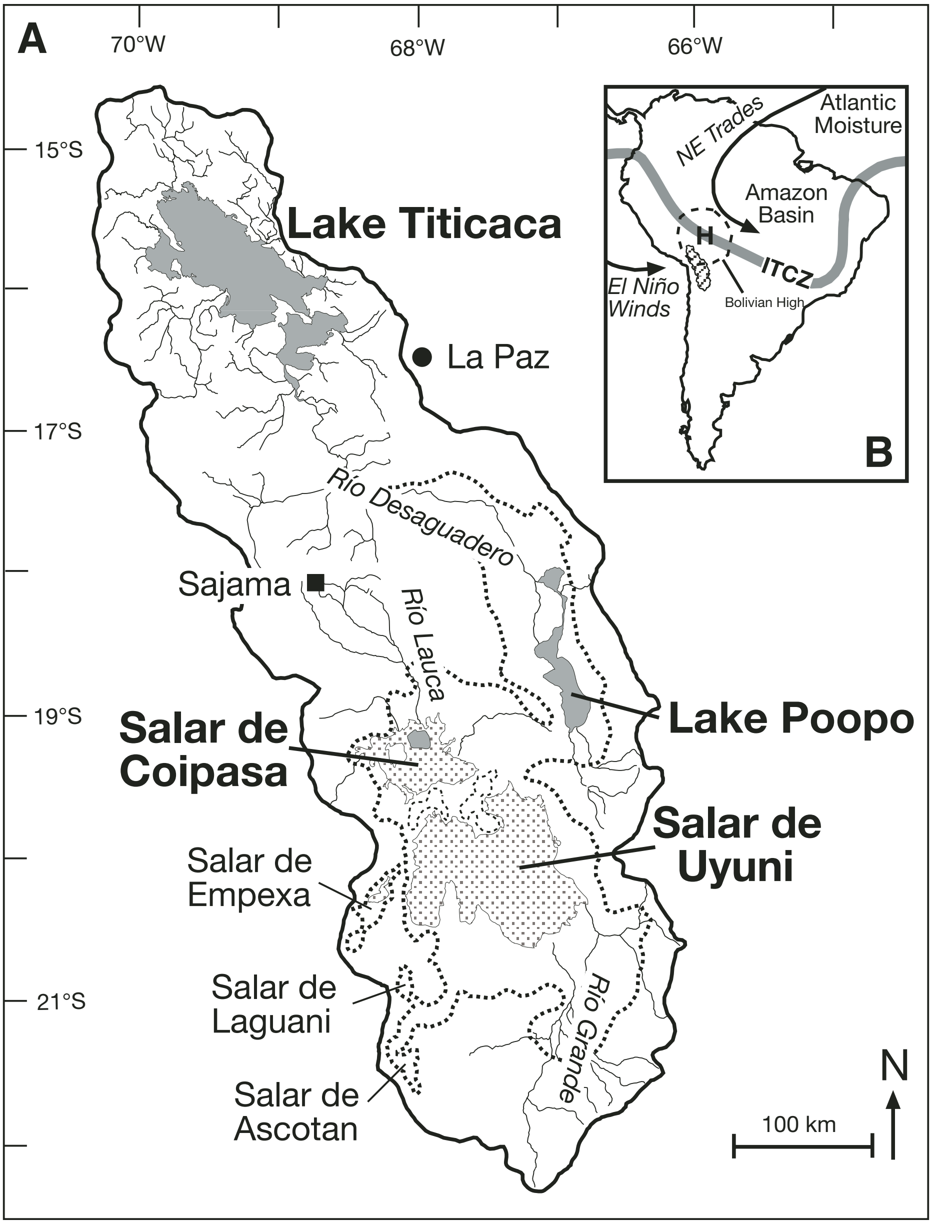

Figure 1. (A) Map of the Titicaca, Poopo, Coipasa, and Uyuni hydrographic basins. Modern lakes are in gray and salars are dotted. The dashed line represents the approximate maximum extent of paleo–Lake Tauca at ~3780 m. (B) The major climate features of South America. The Titicaca, Poopo, Coipasa, and Uyuni drainage basins are striped. ITCZ—Intertropical Convergence Zone.

control on Altiplano moisture balance. Evidence for two deep lakes, generally referred to as the Tauca and Minchin phases, has been studied since before the turn of the century (Minchin, 1882). Several researchers date the Tauca paleolake phase between 19 and 12 ka (Servant and Fontes, 1978; Rondeau, 1990; Bills et al., 1994; Wirrmann and Mourguiart, 1995; Sylvestre et al., 1999; Argollo and Mourguiart, 2000; Fornari et al., 2001), whereas recent evidence from a long core places the Tauca phase between 26 and 14 ka, in-phase with a summer insolation maximum (Baker et al., 2001a). Radiocarbon dates for the Minchin lake phase range from 28 ^{14}C yr B.P. to the limit of ^{14}C dating (Servant and Fontes, 1978; Rondeau, 1990; Baker et al., 2001a).

Here we present a U-Th and ^{14}C chronology spanning 120 k.y. for lake cycles in the Poopo, Uyuni, and Coipasa Basins. Over this time, these basins were twice occupied by some of the largest paleolakes (up to 60,000 km^2) in the Americas. Our new chronology dramatically increases the number of exposures studied, and we replicate many dates using both ^{14}C and U-Th methods. Most critically, a firm U-Th chronology for these basins is required in order to test for possible hard-water effects on ^{14}C dates (Sylvestre et al., 1999) and to date older lakes at or beyond the limit of ^{14}C (Rondeau, 1990). For many stages of the basins' various lake cycles, we are able to closely constrain absolute lake level by dating nearshore or beach deposits.

STUDY AREA

Four very large lake basins (Titicaca, Poopo, Coipasa, and Uyuni) occupy a vast intermontane depression (the Altiplano) spanning 14–22°S (Fig. 1). In the north, Lake Titicaca (3808 m; 8560 km^2) is a deep (>285 m) (Argollo and Mourguiart, 2000), freshwater lake that loses ≤10% of its annual water budget to overflow into the Rio Desaguadero (Roche et al., 1992). The Rio Desaguadero empties into the oligosaline Lake Poopo (3685 m; 2530 km^2), which is separated by the Laka topographic sill (3700 m) from the salars of Coipasa (3656 m) and Uyuni (3653 m). These salt pans occupy 12,100 km^2 (Argollo and Mourguiart, 2000) and in wet years are seasonally filled with shallow water (<4 m) originating primarily from the Rio Lauca and Rio Grande. Paleolakes deep enough to cover the Laka sill at 3700 m integrated the Poopo, Coipasa, and Uyuni Basins, spanned more than 3.5° latitude, and occupied an area of 33,000–60,000 km^2 (Blodgett et al., 1997). The Empexa, Laguani, and Ascotan subbasins extend beyond the Uyuni Basin and were integrated into the deepest lakes at various stages (Fig. 2).

The Bolivian Altiplano lies on the western end of the Amazon rainfall belt, and precipitation occurs primarily during the austral summer. Moisture traverses the Amazon Basin in the summer months when the Intertropical Convergence Zone (ITCZ) is displaced southward and convection is intense in the Amazon (Lenters and Cook, 1997). Deep convection over the Altiplano releases large amounts of energy into the middle and upper troposphere, producing a monsoonal moisture regime (Zhou and Lau, 1998). A moisture source (Fig. 1B) to the north and east of the Altiplano produces a pronounced north-south and east-west precipitation gradient. High spatial variability within any rain event or rainy season is caused by the sporadic nature of the convective afternoon and evening storms that bring rain to the Altiplano (Aceituno and Montecinos, 1993).

The principal control on modern rainfall variability on the Altiplano is the advection of moisture from the eastern lowlands (Vuille and Keimig, 2004) by the trade winds. Variation in

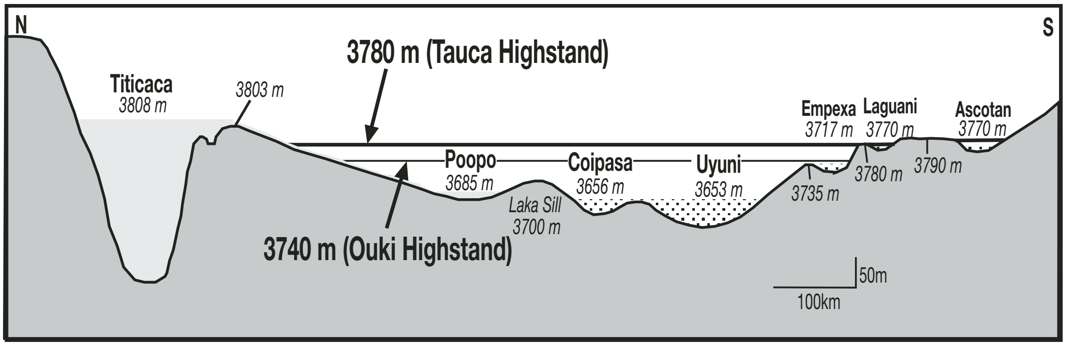

Figure 2. Schematic cross section of the Titicaca, Poopo, Coipasa, and Uyuni hydrographic basins. The elevation of the spillways into the Empexa, Laguani, and Ascotan Basins is known only within ±15 m. Vertical exaggeration is ~830×.

transport of moisture by this means is linked to ENSO, and La Niña years tend to correlate with stronger easterly (trade) winds and more rainfall on the northern Altiplano. Conversely, El Niño years tend to correlate with drought (and westerly wind anomalies) over the Altiplano (Aceituno, 1988; Vuille et al., 1998; Garreaud and Aceituno, 2001; Vuille and Keimig, 2004). Wet years on the northern Altiplano, however, are not necessarily correlated with wet years on the southern Altiplano (Vuille and Keimig, 2004). On the southern Altiplano, the strength and intensity of the Bolivian High (Vuille and Keimig, 2004), which may have extratropical influences in addition to probable links to ENSO, also control interannual precipitation variability (Aceituno and Montecinos, 1993; Lenters and Cook, 1997; Vuille et al., 1998; Vuille and Keimig, 2004).

STRATIGRAPHIC NOMENCLATURE

We use the conceptual framework of alloformations and geosols to describe the stratigraphy of the Poopo, Uyuni, and Coipasa Basins (cf. Morrison, 1978; Oviatt et al., 1994). An alloformation is a "mappable stratiform body of sedimentary rock that is defined and identified on the basis of its bounding discontinuities" (North American Commission on Stratigraphic Nomenclature, 1983, p. 865). The concept of alloformations is most suitable to the reconstruction of lake cycles, because it recognizes that each lake cycle may be represented by diverse lithologic facies (Oviatt et al., 1994). We use the term "geosol" (sensu Morrison, 1978) to describe both buried and surficial soil stratigraphic units that formed between deep-lake cycles.

Where possible, we retain existing nomenclature for the various lake cycles, but the conflicting use of the term "Minchin" prompts us to discontinue the use of this name. Initially, Minchin applied to the deepest lake in the basins (e.g., Steinmann et al., 1904; Ahlfeld and Branisa, 1960). Sediments attributed to Lake Minchin are prominent features in the Poopo Basin and were dated by ^{14}C to ca. 28 ka by Servant and Fontes (1978). More-recent work (e.g., Rondeau, 1990; Fornari et al., 2001; Baker et al., 2001a; Fritz et al., 2004) recognizes two deep-lake cycles in the Poopo, Coipasa, and Uyuni Basins, and generally refers to a lake dating to ca. 45 ka as "Minchin." Thus, the term Minchin has been applied to: (1) the highest-elevation lake in the basin (e.g., Steinmann et al., 1904; Ahlfeld and Branisa, 1960; Bills et al., 1994); (2) sediments exposed and dated at specific localities (e.g., Servant and Fontes, 1978); and (3) a lake dating to ca. 45 ka (e.g., Baker et al., 2001a; Fritz et al., 2004). Since we find that none of these three applications of this term applies to the same lake cycle, we favor dropping this ambiguous term.

METHODS

Field Sampling

Over 30 exposures[1] (Fig. 3) were described and sampled as a part of this study in a comprehensive effort to obtain and replicate records of lake change from multiple localities in all three major basins. Particular effort was directed toward sedimentary deposits associated with the various visible paleoshorelines. This approach to reconstructing lake-level history allows for direct determination of lake level, replication of stratigraphy, and dating by two geochronologic (^{14}C, U-Th) methods. Constructional tufa benches, imbricated gravel deposits, ooid-rich units, and sedimentary successions showing lake transgression or regression provide constraints on absolute paleolake elevation. We sought samples from the bottoms and tops of the lacustrine deposits exposed in dry washes cutting through visible paleoshorelines, ensuring that we repeatedly captured the duration of lake episodes. We addressed the potential incompleteness of any single exposure by replication of stratigraphy at multiple locations. Tufas in the basins have easily identifiable biologic textures (Rouchy et al., 1996), probably indicating formation within the photic zone. Since the absolute depth of tufa formation is not certain, tufas are used to constrain lake elevation only when they are found in association with a constructional bench, are part of a stratigraphic sequence, or formed in a low-light cave.

[1]GSA Data Repository item 2006100, complete site and locality descriptions for every radiometric date and a color copy of Figure 5, is available on the Web at http://www.geosociety.org/pubs/ft2006.htm. Requests may also be sent to editing@geosociety.org.

Laboratory Methods

Radiocarbon Dating

Carbonate samples intended for ^{14}C analyses were treated with 2% H_2O_2 for 2 h to remove organic matter, and reacted with 100% phosphoric acid under vacuum at 50 °C to produce CO_2. Samples of coarse organic material were treated with warm 2*M* HCl, rinsed with deionized water, treated with warm 2% NaOH, and rinsed again with deionized water. The residue (or humin) was dated separately from the base soluble humate fraction. Both humin and humates were combusted at 900 °C in the presence of CuO and silver foil to produce CO_2. Cryogenically purified CO_2 from both organic matter combustion and carbonate hydrolysis was converted to graphite on a pure iron carrier using zinc to catalyze conversion. Radiocarbon analyses of graphite were performed at the National Science Foundation–University of Arizona Accelerator Mass Spectrometry Laboratory. All dates younger than 22 ^{14}C ka were calibrated using CALIB 3.4 (Stuiver and Reimer, 1993) and reported in calendar years relative to 1955 (cal. yr B.P.). One sample dated at 24,000 cal. yr B.P. was converted to calendar years using the quadratic equation of Bard (1998).

U-Th Dating

Most carbonates were dissolved in 2*M* HNO_3 and spiked with mixed ^{233}U-^{229}Th. U and Th were co-precipitated with $FeOH_3$, separated by anion exchange, and measured on a Micromass Sector

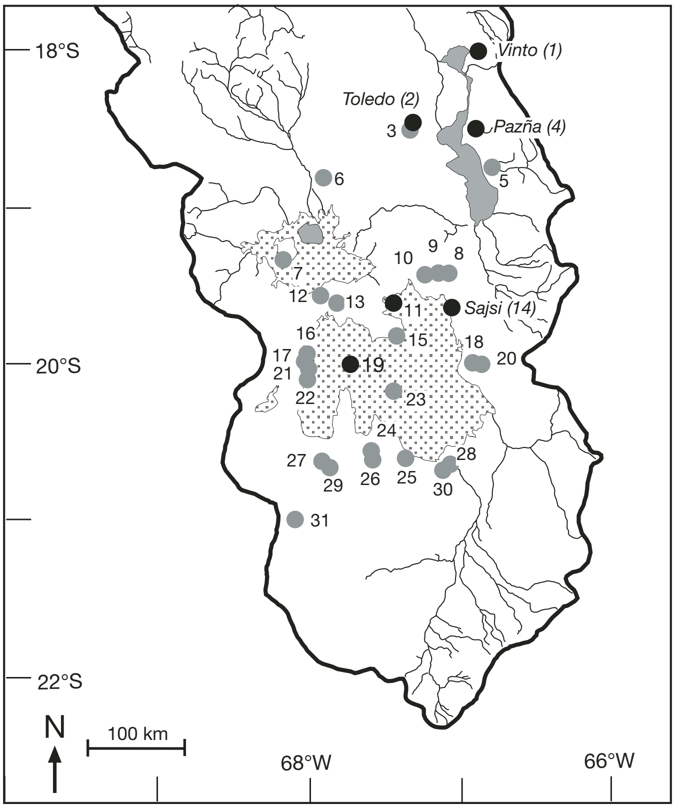

Figure 3. Location of sample sites in the Poopo, Coipasa, and Uyuni Basins. Sites described in the text are represented by black circles. Other symbols are given in Figure 1.

54 thermal ionization mass spectrometer in the Department of Geosciences at the University of Arizona. Details of our analytical procedures are fully discussed in Placzek et al. (2006). By convention, U-Th dates are presented as absolute ages relative to the year of analysis. Thus, an ~50 yr discrepancy exists between the U-Th and calibrated ^{14}C time scales.

DATING CONSIDERATIONS

Radiocarbon Dating

We ^{14}C dated 87 samples (Tables 1 and 2) from the Poopo, Coipasa, and Uyuni Basins. Two potential sources of error in ^{14}C dating of lacustrine carbonates are: (1) contamination by younger carbon, either from secondary organic matter or atmospheric CO_2, which will cause ages to be too young, and (2) ^{14}C reservoir effects, in which lake waters are depleted in ^{14}C and yield ages that are too old. Eighteen paired ^{14}C and U-Th dates allow us to evaluate these two effects (Fig. 4).

For older samples (older than 25 ka), U-Th ages are significantly (more than 50 k.y.) older than associated ^{14}C dates. Hence, U-Th dates from the Ouki Alloformation range from ca. 120 to 98 ka (Table 3), whereas ^{14}C dates from the Ouki Alloformation range between 45.2 and 28.2 ka (Table 2). Large disparities between U-Th and ^{14}C dates from "old" (older than 45 ka) lacustrine tufa are well documented in tufas from the western United States (Lao and Benson, 1988) and East Africa (Hillaire-Marcel et al., 1986), where samples known to be older than 45 ka yield finite ^{14}C ages. Carbonates from these studies contain as much as 3% modern carbon contamination, producing apparent ^{14}C ages as young as 28,700 ^{14}C yr B.P., which is virtually identical to our results, where the percent modern carbon ranges between 0.4% and 3.0% (Table 2).

The reason for this disparity is that as the ^{14}C dating method approaches its limit (ca. 45 ka), small amounts of carbon contamination can produce large shifts in apparent ages. This problem is particularly acute in carbonates, even where shells retain their original aragonite composition (Brennan and Quade, 1997). The cause for this must be related to the surface exchange of atmospheric CO_2 with carbonate radicals, a process observed when carbonates gradually take up CO_2 after grinding (Samos, 1949). Thus, we view all ^{14}C dates on carbonate older than 28,000 ^{14}C yr B.P. as minimum ages.

By contrast, U-Th and ^{14}C dates from the younger (younger than 18 ka) part of the record match very well (Fig. 4). This strong correlation suggests that (1) surface sorption of CO_2 by younger aragonite shell and tufa is minimal, and (2) that ^{14}C reservoir effects in Lake Tauca are small, if present at all. Additionally, ~3% contamination, equivalent to the most extreme cases from Ouki Alloformation carbonates, would reduce the age of ca. 16 ka samples by ~1500 yr, similar to the overall uncertainty in calibrated ages for this time. In our view, this example is a worst-case scenario, since Ouki-age carbonates are an order of magnitude older than Tauca-age carbonates, and were inundated by the younger Tauca deep-lake cycle, allowing greater opportunity for contamination by younger carbon.

Radiocarbon deficiencies in excess of 10,000 yr have been noted in modern waters from the Altiplano (Geyh et al., 1999); therefore, verification of our ^{14}C-based chronology is required. The agreement between paired ^{14}C and U-Th dates from samples dating between 18 and 11 ka argues against ^{14}C reservoir effects over this time (Fig. 4). We also obtained nine ^{14}C dates from semi-aquatic snails, family Succineidae, which is a taxon known to provide reliable ages in locations with ^{14}C reservoir effects (Pigati et al., 2004).

U-Th Dating

Our lake-level reconstruction relies on dates with small corrections for initial ^{230}Th associated with siliciclastic material incorporated in the carbonate. Initial ^{230}Th daughter was tracked by ^{232}Th, and a correction was made using an initial $^{230}Th/^{232}Th$ ratio derived from isochron plots (Placzek et al., 2006). Reported age uncertainties include propagated 2σ envelopes on isotope ratios decay constants, as well as error on our assumed initial $^{230}Th/^{232}Th$ ratio. Age uncertainties are principally a function of the amount of initial Th incorporated into a carbonate. Here we present dates with uncertainties that are less than half the duration of the relevant lake cycle, and such dates are precise enough to be stratigraphically meaningful. The entire population of U-Th dates (>90) from the Bolivian Altiplano is, however, consistent with the chronology framed by these dates, and a detailed discussion of U-Th dating of Altiplano carbonates is presented elsewhere (Placzek et al., 2006).

STRATIGRAPHY AND CHRONOLOGY

In this section, we present our stratigraphic framework and chronologic evidence for

TABLE 1. RADIOCARBON DATA AND AGES

Sample number (Lab ID)	^{14}C age (yr)	Age (cal. yr B.P.)[†]	Error (+)	Error (–)	Altitude (m)	Uncertainty (m)[‡]	Site number	Material[§]
Holocene								
54462 (S-21–3)	529 ± 34	530	90	20	3667	5	14L	Black mat
54440 (S-21–2)	1833 ± 48	1710	150	140	3667	5	14L	Succineidae
54415 (C-21–2A)	2286 ± 37	2330	40	180	3705	10	8A	Terrestrial snail
49075 (ColK-8)	3028 ± 46	3180	170	180	3785	10	26	Black mat
54456 (C-11–6B)	5444 ± 47	6240	60	240	3660	5	7	Charcoal
54457 (C-11–6A)	5591 ± 45	6370	70	90	3660	5	7	Charcoal
54430 (P-3–4)	7641 ± 59	8400	130	70	3712	10	9A	Terrestrial snail
49074 (ColK-4)	9413 ± 52	10,610	130	190	3785	10	26	Black mat
Coipasa Lake Cycle								
54421 (S-2–1)	9564 ± 61	11,040	120	440	3660	5	14F	Succineidae
54451 (U-9–6)[#]	9913 ± 83	11,260	430	80	3703	20	25	Tufa
60473 (U-9–6 duplicate)	9987 ± 60	11,300	610	100	3703	20	25	Tufa
54464 (U-27a-6)	9938 ± 56	11,260	300	60	3657	5	19	Tufa
60468 (S-6–3)	10,036 ± 55	11,410	530	170	3660	5	^{14}C	Tufa
58648 (S-7–2)	10,295 ± 55	12,090	640	390	3658	5	14B	Succineidae
58651 (S-3–3)	10,336 ± 54	12,290	500	520	3662	5	14E	Succineidae
58650 (S-4–1)	10,394 ± 58	12,320	492	541	3661	5	14D	Succineidae
54429 (S-6–3a)	10,672 ± 62	12,720	240	370	3660	5	14C	Succineidae
58658 (U-31–4)	10,623 ± 55	12,800	140	470	3655	5	23	Tufa
54447 (U-10)	11,031 ± 62	13,000	170	340	3662	10	30	Tufa
49061 (B-17)[#]	11,088 ± 54	13,100	80	430	3695	5	15	Tufa
58656 (U-31–3)	11,582 ± 65	13,480	360	290	3655	5	23	Tufa
Tauca Lake Cycle								
Regression								
54431 (P-11–2)	11,312 ± 61	13,180	600	160	3725	10	3A	*Littoridina*
49060 (B-14)[#]	11,944 ± 61	13,940	1330	310	3735	5	15	Tufa
60472 (C-3–10)	12,088 ± 81	14,090	1270	420	3712	10	6	*Littoridina*
58654 (S-20–1)	12,140 ± 220	14,100	1340	580	3666	5	14K	Gravel
49070 (P-25)	12,147 ± 67	14,110	1270	410	3719	10	4A	*Littoridina*
54425 (C-22–1)	12,190 ± 67	14,120	1280	290	3705	10	8B	*Littoridina*
54411 (C-24–1)	12,266 ± 66	14,240	1200	380	3659	10	12	*Littoridina*
49078 (P-3–9)	12,292 ± 62	14,270	1180	420	3738	10	9B	*Littoridina*
49059 (B-12)[#]	12,321 ± 61	14,290	1170	430	3740	5	15	Tufa
49066 (P-24)	12,328 ± 59	14,290	1170	430	3717	10	4A	*Littoridina*
49065 (P-3–2)	12,446 ± 60	14,350	1170	210	3710	10	9A	*Littoridina*
Highstand								
49079 (U-8–1)	12,193 ± 58	14,100	1280	290	3761	10	27	*Littoridina*
49072 (U-2–1)	12,311 ± 59	14,280	1170	420	3763	10	15	*Littoridina*
49071 (P-4–4)	12,397 ± 63	14,330	1170	210	3745	10	10A	*Littoridina*
54434 (C-2–3)	12,448 ± 85	14,350	1180	220	3770	10	13	*Littoridina*
49067 (U-5u)	12,596 ± 60	14,640	940	440	3772	10	22	*Littoridina*
49069 (Chita-9)	12,598 ± 60	14,640	560	440	3715	10	20B	*Littoridina*
54419 (S-19–2)	12,482 ± 68	14,970	560	830	3665	5	14K	*Littoridina*
54433 (P-7–6)	12,484 ± 66	14,980	550	840	3720	10	1B	*Littoridina*
54417 (S-2–3)	12,574 ± 77	15,170	410	990	3661	5	14F	*Littoridina*
54441 (P-3–1)	12,625 ± 68	15,230	390	1010	3710	10	9A	Carbonate sediments
49063 (Chita-4)	12,684 ± 68	15,300	390	1020	3725	10	20A	*Littoridina*
49064 (Chita-10)	12,685 ± 60	15,300	390	1020	3715	10	20B	*Littoridina*
49068 (U-12–2)	12,733 ± 72	15,350	410	1040	3713	10	20C	*Littoridina*
49081 (U-13–2)	12,760 ± 62	15,380	400	1040	3727	10	14M	*Littoridina*
54416 (C-21–2B)	12,770 ± 100	15,390	440	1060	3705	10	8B	*Littoridina*
54412 (S-19–0)	12,774 ± 99	15,390	440	1060	3665	5	14K	*Littoridina*
49076 (P-4–1)	12,820 ± 65	15,440	420	1050	3738	10	10B	*Littoridina*
54420 (RG2–2)	12,846 ± 70	15,460	430	1490	3667	10	28	*Littoridina*
58653 (S-14–2)	12,917 ± 67	15,520	450	1050	3666	5	14J	*Littoridina*
54463 (U-27a-5)[#]	12,925 ± 93	15,530	480	1080	3657	5	19	Tufa
54446 (U-7–1)[#]	12,942 ± 70	15,550	470	1060	3770	10	29	Tufa
54422 (P-8–1)	13,168 ± 74	15,800	480	1070	3725	10	1A	*Littoridina*
58652 (S-7–1)	13,176 ± 66	15,810	475	1060	3658	5	14B	*Littoridina*
54432 (Chita ash)	13,199 ± 84	15,840	490	1070	3725	10	20A	*Littoridina*
49080 (U-8–4)	13,329 ± 68	15,990	480	950	3763	10	27	*Littoridina*
54418 (Chig1–1)	13,494 ± 73	16,180	490	460	3760	20	31	*Littoridina*
54442 (C-2–2)	13,529 ± 72	16,220	490	460	3770	10	13	Carbonate sediments
49062 (Chita-2)	13,577 ± 66	16,280	490	460	3723	10	20A	*Characeae*
54427 (P-7–7)	13,660 ± 100	16,370	520	490	3720	10	1C	*Littoridina*
54438 (P-12–1)	13,661 ± 73	16,370	500	460	3750	10	3B	*Littoridina*
Transgression								
54444 (U-4–1)[#]	14,231 ± 76	17,030	530	490	3700	20	17	Tufa
54448 (U-29–1)	14,350 ± 110	17,170	570	520	3670	10	18	Ooids
60471 (U-27a-3)	14,375 ± 73	17,190	540	490	3659	5	19	Tufa
58660 (S-17–5)	14,684 ± 73	17,550	560	500	3665	5	14G	Tufa
60470 (S-17–6)	14,767 ± 73	17,640	570	510	3666	5	14G	Tufa
54460 (U-27a-2)[#]	14,941 ± 80	17,850	590	520	3657	5	19	Tufa
60469 (S-6–1)	15,044 ± 88	17,960	600	540	3659	5	14C	Tufa
58649 (S-3–1)	18,800 ± 1100	22,290	2530	2600	3662	5	14E	Tufa
Sajsi Lake Cycle								
58657 (S-17–7)	17,080 ± 720	20,310	1790	1750	3662	5	14	Tufa
58659 (S-17–4)	17,255 ± 93	20,510	690	660	3663	5	14G	Tufa
54436 (S-17–2)	20,090 ± 130	23,770	780	790	3661	5	14G	Ooids
54437 (S-17–1)	20,830 ± 140	24,000[††]			3660	5	14G	Ooids

[†]Calibrated with CALIB 4.3 using the probabilities method. Southern Hemisphere correction of 24 yr was used.
[‡]Uncertainty in sample-site elevations varies according to the method of elevation determination.
[§]All tufas are calcite and all shells are aragonite.
[#]Paired U-Th dates are available in Table 2.
[††]Calibrated using the quadratic equation of Bard (1998).

TABLE 2. RADIOCARBON DATA AND AGES FOR PRE-TAUCA LAKE CYCLES

Sample number	¹⁴C age (yr)†	Modern C (%)	Site no.	Material	Lake cycle	U-Th sample no.	U-Th date (yr)†
54414	28,170 ± 290	3.00	8A	*Littoridina*	Ouki		
54461	28,500 ± 1900	2.88	4B	Wood (humates)‡	Ouki	Pazna-9	110,770 ± 1760
54426	30,950 ± 630	2.12	8B	*Littoridina*	Ouki		
54424	31,490 ± 500	1.98	8B	*Littoridina*	Ouki		
54413	31,930 ± 450	1.88	1B	*Littoridina*	Ouki	P-8–2	104,110 ± 730
49077	34,110 ± 570	1.43	4A	*Littoridina*	Ouki	P-17	106,040 ± 1040
54450	34,300 ± 650	1.40	2	Tufa	Ouki	B-3	119,720 ± 3020
54445	36,930 ± 830	1.01	1B	*Littoridina*	Ouki	P-8–2	104,110 ± 730
54458	37,270 ± 880	0.97	4B	Wood (humin)‡	Ouki	Pazna-9	110,770 ± 1760
54439	37,930 ± 990	0.89	4B	Bulimulidae	Ouki	Pazna-9	110,770 ± 1760
54459	40,300 ± 1200	0.66	19	Tufa	Inca Huasi	U-28–1	45,760 ± 440
54435	40,600 ± 1200	0.64	1A	Tufa	Ouki	P-8–2	104,110 ± 730
54449	45,200 ± 2300	0.36	5	Tufa	Ouki	P-5–3	102,620 ± 860

†We view all ^{14}C dates on carbonate older than 28,000 ^{14}C yr B.P. as minimum ages and present such dates as uncalibrated ^{14}C yr B.P.
‡Different fractions from the same wood sample.

lake-level changes, building upon the earlier efforts of Servant and Fontes (1978), Lavenu et al. (1984), Wirrmann and Mourguiart (1995), and Sylvestre et al. (1999). We will describe a few key localities that typify the stratigraphic succession and ages of a much broader spectrum of sites compiled in the GSA repository (see footnote one). Lacustrine sediments from the two deep-lake cycles and represented by the Ouki (~3740 m; ca. 120–98 ka) and Tauca (~3770 m; 18.1–14.1 ka) Alloformations are widely exposed. Between the Ouki and Tauca deep-lake cycles, the lake remained below 3700 m. Above that elevation, deep pedogenesis (the Toledo Geosol) and minor alluviation occurred between 98 and 18.1 ka. The weakly developed Hanco Geosol caps the Tauca Alloformation and formed over the past ca. 12 k.y. Minor lake cycles include: Salinas (95–80 ka), Inca Huasi (ca. 46 ka), Sajsi (24–20.5 ka), and Coipasa (13–11 ka).

Ouki Alloformation

The Vinto site (Fig. 3) in the northeastern Poopo Basin contains many exposures of the newly identified Ouki Alloformation, and is also where carbonates originally attributed to “Lake Minchin” were described (Servant and Fontes, 1978; Wirrmann and Mourguiart, 1995; Rouchy et al., 1996). The deposits are mainly observed in carbonate quarries used for stucco production at 3720–3740 m, just below the elevation of the highest deposits of the Ouki Alloformation. The tufas at Vinto, like other tufas of the Ouki Alloformation, have a distinct morphology, consisting of large inverted cones of relatively porous carbonate up to 3.5 m high and several meters in diameter, with a dense, mammillary exterior crust (Fig. 5A). Material from the interior of a tufa head at Vinto yielded a U-Th date of 104,110 ± 730 (Fig. 6; Table 2). Coarse calcare-

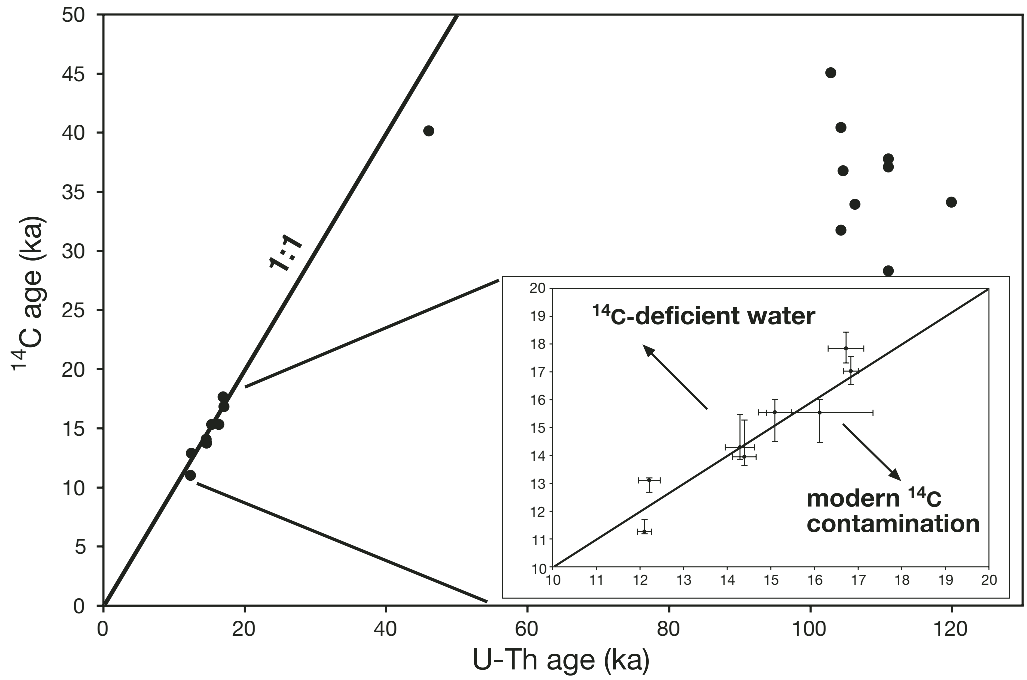

Figure 4. U-Th and ^{14}C dates. The younger dates (younger than 50 ka) are paired samples from the same tufas, and older dates come from the same stratigraphic horizon (Table 2). Samples with U-Th ages older than 95 ka yield ^{14}C ages from 28 to 45 ka. We interpret these as infinite ^{14}C ages contaminated with variable small amounts of younger carbon. The inset graph details the period 10–20 ka; note agreement for ^{14}C and U-Th ages for the 10–20 ka results. By convention, ^{14}C dates are relative to 1955, but U-Th dates are relative to the year of analysis. In this figure, U-Th dates younger than 20 ka are converted to years before 1955 (cal. yr B.P.). Arrows indicate the potential effects of contamination by ^{14}C-enriched (modern) and ^{14}C-deficient (ancient) water on ^{14}C dates.

ous sand fills areas between the large tufa heads. These sands include bedded or cross-bedded coquinas containing abundant aquatic snails (*Littoridina* sp. and *Biomphalaria andecola*) and bivalves (*Pisidium* sp.). Shells and tufa from this alloformation yield ^{14}C dates between 31,930 and 40,600 ^{14}C yr B.P. (Table 2). This is consistent with previous ^{14}C determinations from these deposits (Rondeau, 1990; Wirrmann and Mourguiart, 1995). As discussed in the previous section on dating, we regard these ^{14}C dates as spuriously young.

The Ouki Alloformation is also well exposed just north of Pazña (Fig. 3) in the northeastern Poopo Basin. The Ouki Alloformation consists of coarse calcareous sands with abundant aquatic snails (*Littoridina* [two species], and *Biomphalaria andecola*), *Pisidium* sp., ostracodes, and small (<10 cm) bulbous nodules of needle-like tufa. These nodules are in growth positions near the base of the Ouki Alloformation and yield two U-Th dates that almost overlap within error (109,100 ± 1870 and 106,040 ± 1040) (Fig. 7; Table 3). *Littoridina* sp. shells from the same stratigraphic level in this unit, however, yield a ^{14}C date of 34,110 ± 570 ^{14}C yr B.P.; again we reject this date as slightly contaminated. Nearby, quarrying exposes the large inverted tufa cones typical of the Ouki Alloformation. Dates from these cones are: 110,770 ± 1760 at the base, 102,810 ± 3020 in the center of the same cone, and 104,940 ± 3160 from the dense mammillary crust that forms the exterior of the tufa heads.

At the Pazña locality, diverse material of known "infinite" ^{14}C age was selected for further investigation into the contamination of organic material by younger carbon. A Bulimulidae (terrestrial) snail returned a date of 37,930 ± 990 ^{14}C yr B.P. (Table 3), and carbonized wood dated to 37,270 ± 880 ^{14}C yr B.P. The humate fraction (base soluble) of this wood returned a date of 28,500 ± 1900 ^{14}C yr B.P. Thus, wood in the Ouki Alloformation (Table 2) is also likely contaminated, a situation viewed to be much more widespread in old ^{14}C samples than previously thought (Bird et al., 1999).

The oldest and youngest dates from the Ouki Alloformation come from lower elevation sites but have relatively large (>3 k.y.) uncertainties due to the incorporation of significant initial Th. Many of these dates overlap, within error, the more precise dates from Vinto and Pazña. Several dates are from exposures near the settlement of Toledo on the west side of the Poopo Basin (Fig. 3; ~3715 m). At site 5 (Fig. 3; ~3700 m), several dates are available from the same tufa head, including one precise age of 102,620 ± 860. Lake Ouki transgression and regression cycles are, thus, not well constrained, as U-Th dates from lower elevations have relatively high errors.

TABLE 3. U-Th DATA AND AGES

Sample	$^{230}Th/^{234}U$ age (yr)	2σ (yr)	$^{230}Th/^{232}Th$ (activity)	$^{234}U/^{238}U_{initial}$ (activity)	Altitude (m)‡	Site number
Coipasa Lake Cycle						
U-9–6†	12,150	160	51.7	1.76000	3700 ± 25	25
B-17†	12,260	250	36.4	1.79000	3695 ± 5	15
B-17 duplicate	12,260	330	21.0	1.77000	3695 ± 5	15
U-27a-8	12,530	200	32.4	1.76000	3669 ± 5	19
S-10–3	11,970	1290	4.0	1.92000	3653 ± 5	14
RG 1–3	12,540	1080	5.2	1.78000	3667 ± 10	28
U-27a-6i	13,310	1090	4.8	1.74000	3657 ± 5	19
Tauca Lake Cycle						
Regression						
B-14†	14,440	340	29.4	1.63000	3735 ± 5	15
B-12†	14,340	270	22.8	1.73000	3740 ± 5	15
Highstand						
B-6	13,320	960	9.2	1.53000	3768 ± 5	15
B-8	14,560	660	13.4	1.64000	3763 ± 5	15
B-8 duplicate	15,970	1160	5.1	1.60000	3763 ± 5	15
U-7–1†	15,140	300	22.1	1.62000	3770 ± 10	29
U-5–2	14,720	660	8.4	1.61000	3769 ± 10	21
U-9–2	15,490	600	9.3	1.61000	3760 ± 10	25
U-22–5	16,290	270	61.7	1.59000	3665 ± 10	11
U-27a-5†	16,170	1220	11.5	1.63000	3657 ± 5	19
B-9§	15,540	490	12.6	1.60000	3765 ± 5	15
B-9a	15,390	540	11.3	1.61000	3765 ± 5	15
B-9ab	14,030	860	6.9	1.60000	3765 ± 5	15
B-9ac	15,590	720	8.6	1.60000	3765 ± 5	15
B-9e	12,230	1220	11.5	1.63000	3765 ± 5	15
B-9d	15,660	1440	4.8	1.59000	3765 ± 5	15
Transgression						
U-27a-2†	16,770	410	17.2	1.59000	3657 ± 5	19
U-4–1†	16,880	170	95.1	1.58000	3700 ± 20	17
U-5–3	17,950	900	7.5	1.57000	3695 ± 10	21
Inca Huasi Lake Cycle						
U-28–1	45,760	440	82.6	1.58000	3661 ± 5	16
U-27a-1	46,330	660	26.8	1.60000	3657 ± 5	19
U-27a-1 duplicate	47,160	860	25.5	1.57000	3657 ± 5	19
Salinas Lake Cycle						
U-31–1	80,460	1670	26.7	1.66000	3657 ± 5	23
U-22–3	86,320	5220	38.4	1.64000	3660 ± 10	11
U-26–3	88,680	1090	117.1	1.67000	3667 ± 5	24
U-22–1	88,010	9790	41.2	1.65000	3658 ± 5	11
U-26–2	90,210	2100	40.3	1.68000	3665 ± 5	24
U-22–2	96,020	2910	72.4	1.69000	3660 ± 5	11
Ouki Lake Cycle						
P-5–1	96,740	5560	4.9	1.57000	3698 ± 5	5
P-5–1 duplicate	98,710	6190	25.8	1.59000	3698 ± 5	5
P-5–2	101,280	2240	15.1	1.58000	3700 ± 5	5
P-5–5	100,200	8370	3.7	1.56000	3704 ± 5	5
P-5–3	102,620	860	55.6	1.60000	3702 ± 5	5
P-11	102,810	3020	10.2	1.56000	3715 ± 5	4B
P-8–2	104,110	730	107.2	1.61000	3728 ± 5	1B
Pazna 2–4	104,940	3160	14.4	1.57000	3715 ± 5	4B
P-17	106,040	1040	69.4	1.54000	3718 ± 5	4A
P-17 duplicate	109,100	1870	58.7	1.56000	3718 ± 5	4A
Pazna-9	110,770	1760	35.3	1.63000	3714 ± 5	4B
T-23	110,470	7810	3.8	1.57000	3710 ± 10	2
T-1	113,290	7820	4.2	1.62000	3708 ± 10	2
T-3	114,180	5440	5.4	1.58000	3709 ± 10	2
P-5–4	113,720	4540	7.5	1.58000	3702 ± 5	5
B-3	119,720	3020	14.7	1.68000	3711 ± 10	2
T-9	119,390	3490	12.3	1.66000	3714 ± 10	2
T-13	120,290	6590	5.2	1.64000	3715 ± 10	2
T-12	125,990	9580	4.5	1.66000	3717 ± 10	2

†Paired ^{14}C dates are available in Table 1.

‡Uncertainty in sample-site elevations varies according to the method of elevation determination.

§Repeat analyses of sample B-9 is described in Placzek et al. (2006).

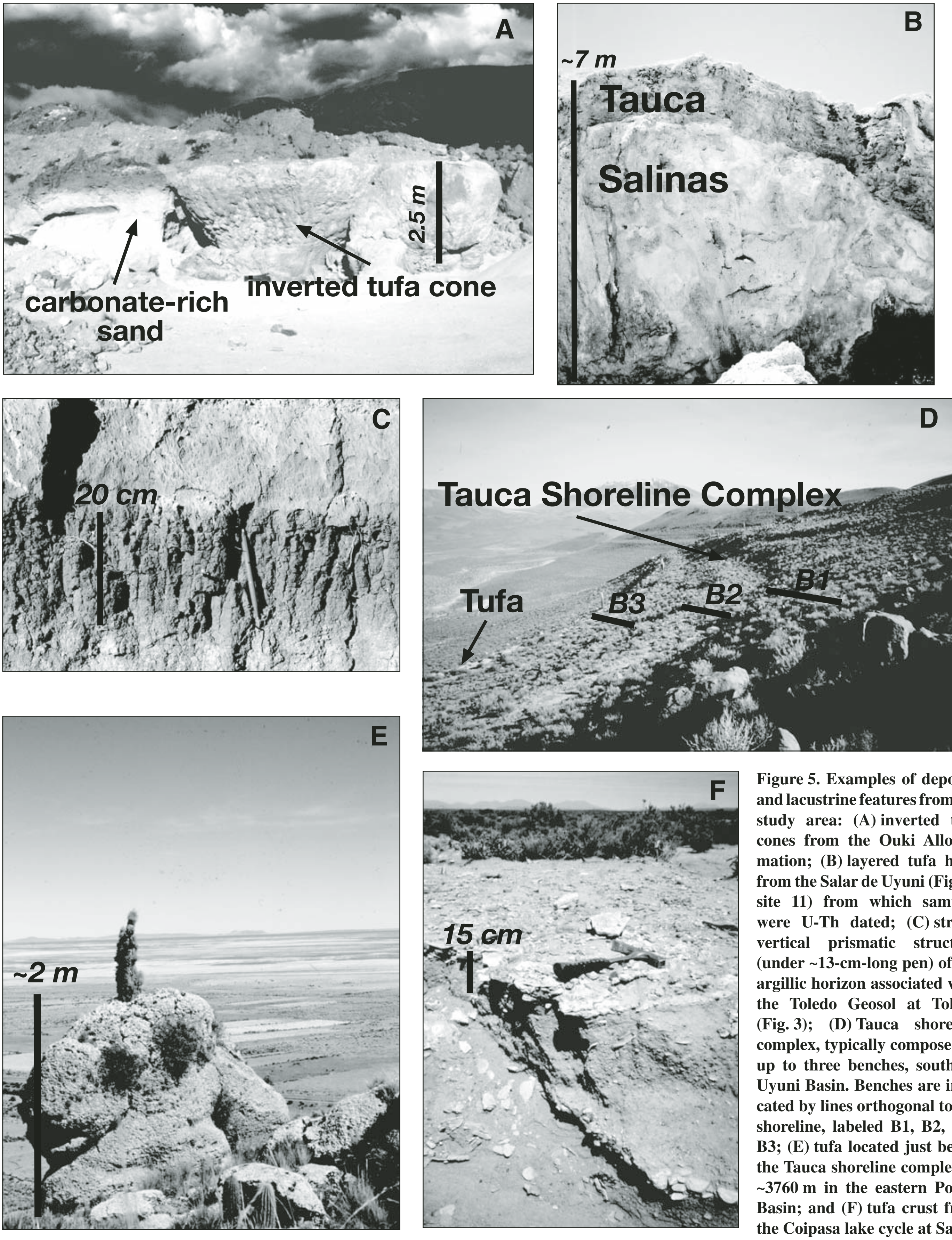

Figure 5. Examples of deposits and lacustrine features from the study area: (A) inverted tufa cones from the Ouki Alloformation; (B) layered tufa head from the Salar de Uyuni (Fig. 3; site 11) from which samples were U-Th dated; (C) strong vertical prismatic structure (under ~13-cm-long pen) of the argillic horizon associated with the Toledo Geosol at Toledo (Fig. 3); (D) Tauca shoreline complex, typically composed of up to three benches, southern Uyuni Basin. Benches are indicated by lines orthogonal to the shoreline, labeled B1, B2, and B3; (E) tufa located just below the Tauca shoreline complex at ~3760 m in the eastern Poopo Basin; and (F) tufa crust from the Coipasa lake cycle at Sajsi.

The U-Th dates from Vinto and Pazña provide critical constraints on the age of the base of the Ouki Alloformation near the Ouki highstand (Fig. 8) and indicate that Lake Ouki persisted at or slightly above 3720 m from ca. 115 to 100 ka. Although no actual shoreline bench of Ouki age is visible, we interpret the tufas and coquinas exposed at these localities as nearshore deposits. *Biomphalaria andecola*, in particular, prefers shallow water depths ≤1 m (Malek, 1985). From this we reconstruct a minimum highstand elevation for the Ouki of ~3735 m. We have not observed the Ouki Alloformation cropping out above 3740 m, which constrains the maximum depth of Lake Ouki to ~80 m.

The Salinas Lake Cycle (<3670 m)

There is evidence from tufas for a lake stand at very low elevation between ca. 95 and 80 ka. Very large mounds (up to 7 m high) of tufa with conical shapes (Fig. 5B) are exposed at low elevation (≤15 m above the modern salar) around the Uyuni Basin, and date by U-Th (Table 2) to 96,020 ± 2910, 88,010 ± 9790, and 86,320 ± 5220 (Fig. 3; site 11); 80,460 ± 1670 (Fig. 3; site 23); and 90,210 ± 2100 and 88,680 ± 1090 (Fig. 3; site 24). The oldest of these dates overlaps, within error, with the youngest dates from the Ouki Alloformation (Table 3). Dates between 90 and 100 ka have large errors, permitting two interpretations of the lake history during this period. One interpretation is that the 100–80 ka tufas date the waning stages of Lake Ouki. The other view is that Lake Ouki dried up and was followed by a modest expansion between 90 and 80 ka. We tentatively attribute these tufas to the Salinas lake cycle, but recognize that this “lake cycle”

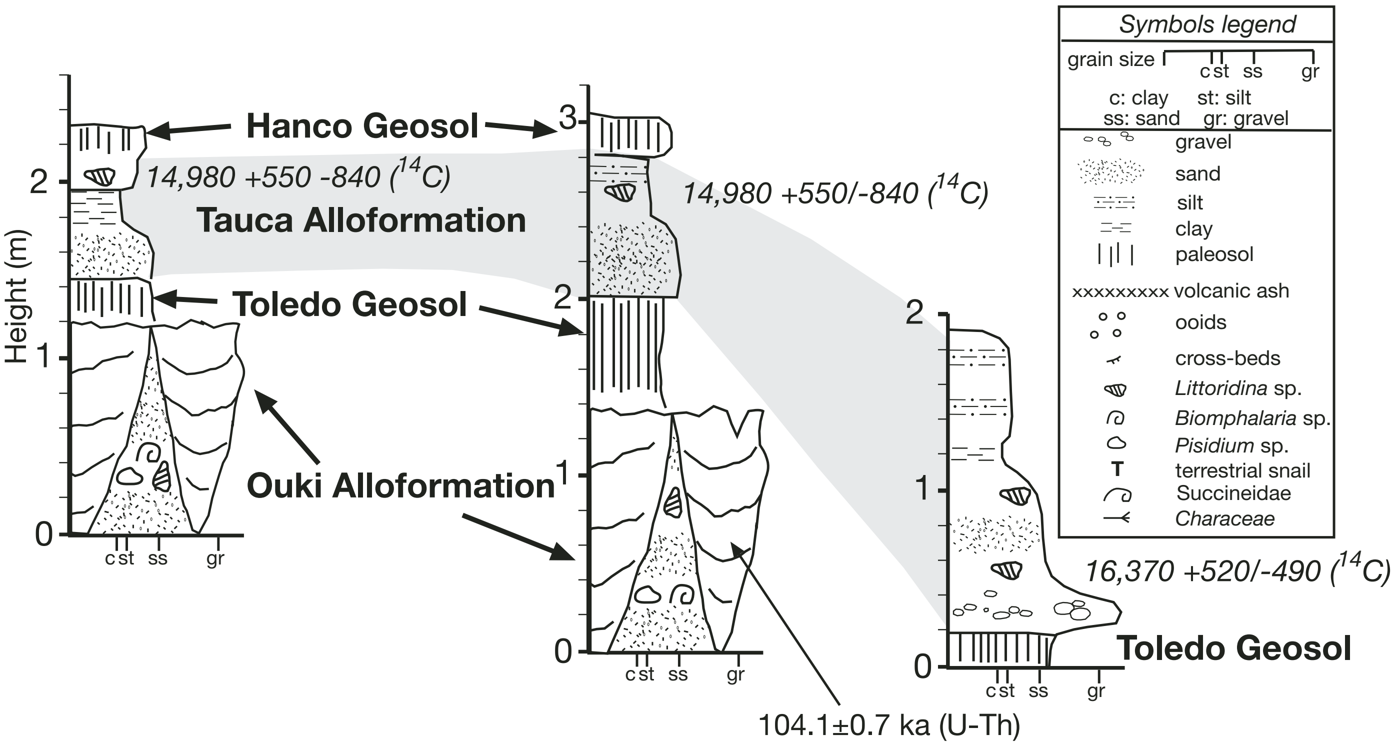

Figure 6. The stratigraphy exposed at Vinto in the northeast Poopo Basin (Fig. 3). Sections are separated from each other by no more than 500 m and were acquired within an area quarried for carbonate.

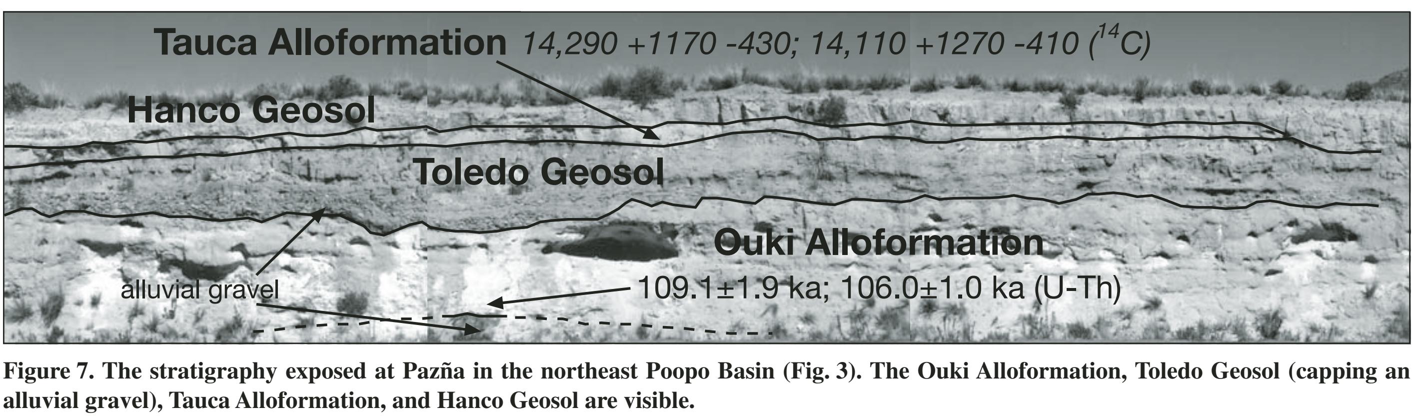

Figure 7. The stratigraphy exposed at Pazña in the northeast Poopo Basin (Fig. 3). The Ouki Alloformation, Toledo Geosol (capping an alluvial gravel), Tauca Alloformation, and Hanco Geosol are visible.

may simply be the waning phases of the Ouki lake cycle.

Toledo Geosol (>3700 m)

The Toledo Geosol is a well-developed paleosol that divides the Ouki and Tauca Alloformations at several sites (e.g., Vinto, Pazña, and Toledo) at elevations above ~3700 m. At Toledo (Fig. 3), this geosol consists of a strikingly reddish (5YR 5/4–3/4 dry), clayey argillic horizon with prismatic peds and visible clay skins (Fig. 5C). A stage II calcic horizon (sensu Gile et al., 1966) occurs beneath the argillic horizon. At some localities, the Toledo Geosol is a simple profile (Fig. 5C), whereas at other locations it consists of stacked paleosols separated by thin layers of alluvium.

The strong development of the Toledo Geosol attests to a prolonged period of stability, and hence no lake transgressions above 3700 m between 95 and 17 ka. The strong soil development also qualitatively supports the long duration of exposure (~80 k.y.) indicated by the U-Th dates. The duration of exposure (~10 k.y.) implied by ^{14}C dates from the Ouki Alloformation is highly implausible, especially given the much stronger development of the Toledo Geosol compared to the Hanco Geosol (see following), which formed over the last ~12 k.y.

The Inca Huasi Lake Cycle (<3670 m)

Three U-Th dates from tufas exposed within meters of the present Salar de Uyuni attest to a minor lake cycle that we designate as the Inca Huasi. The dense interior portion of a tufa in the western margin of Salar de Uyuni (Fig. 3; site 16) U-Th dates to 45,760 ± 440, and a layered bioherm from an island (Fig. 3; site 19) dates to 46,330 ± 660, with a replicate date of 47,120 ± 860 (Fig. 9). In both cases, these tufas are encased in Lake Tauca–age tufa layers. The dense texture of the Inca Huasi tufas separates these tufas from younger encrustations (Fig. 9).

We conclude that the lake level rose around 46 ka to perhaps 10 m depth. This lake cycle is not observed in the exposed shoreline stratigraphy, so we cannot be more specific concerning lake depth. Moreover, tufas of this age are missing from other sequences at low elevation, perhaps because this was a short and shallow lake cycle.

The Sajsi Lake Cycle (≤3670 m)

A minor pre-Tauca lake cycle is exposed in stratigraphy along the Rio Sajsi, ~7 m above the modern salar on the east side of the Uyuni Basin (Fig. 3). Here, continuous low-elevation exposures along cutbanks allowed a detailed examination of the Sajsi, Tauca, and Coipasa lake cycles (Fig. 10). The oldest lake deposits in these exposures are carbonate-cemented, cross-bedded ooids that yield a ^{14}C date of 24,000 cal. yr B.P., overlain by one meter of loosely consolidated ooids that date by ^{14}C to 23,770 +780/–790 cal. yr B.P. (Fig. 10G). Arborescent tufa heads (^{14}C: 20,510 +690/–660 cal. yr B.P.) rest on the ooid beds, and very large (>2 m diameter) tufa heads found nearby ^{14}C date to 20,310 +1790/–1750 cal. yr B.P. Oolitic sands occur between the tufa heads (Fig. 11H), and a reddish (5YR 5/4), very poorly sorted, subangular sand overlies the ca. 20.5 ka tufa (Figs. 10G, 10H, and 10I).

We view this sequence as representing a shallow (3670 m or <17 m deep) lake cycle (the "Sajsi") over the period 24 to ca. 20 ka. The transgression of the lake is clearly indicated by oolitic beach deposits grading upward into shallow-water tufas mixed with more oolitic beach sand. The overlying poorly sorted, reddish sands are identical in appearance to modern alluvium in Sajsi wash. We take this as evidence that the lake dropped below this local elevation of 3665 m after 20.5 ka.

Tauca Alloformation

Tauca Lake-Cycle Transgression

The transgression of Lake Tauca can be seen in buried stratigraphy and a succession of surface tufas at <3700 m. Chronologic control on this transgression comes principally from the Uyuni Basin. A layered tufa (Fig. 3; site 19) ^{14}C dates to 17,850 +590/–520 cal. yr B.P., where it is cemented directly onto an older tufa belonging to the Inca Huasi lake cycle (Fig. 9). A succeeding layer from the same tufa yielded overlapping dates of 16,770 ± 410 (U-Th) and 17,190 +550/–490 cal. yr B.P. (^{14}C). A thin, buried carbonate crust (^{14}C: 17,960 +600/–540 cal. yr B.P.) marks the transgression of Lake Tauca at Sajsi (Fig. 10C; 3657 m). At a slightly higher elevation of 3663 m at Sajsi (Fig. 10G), dense mammillary tufa heads (^{14}C: 17,550 +560/–500 and 17,640 +570/–510 cal. yr B.P.) resting on fluvial sands mark the base of the Tauca Alloformation.

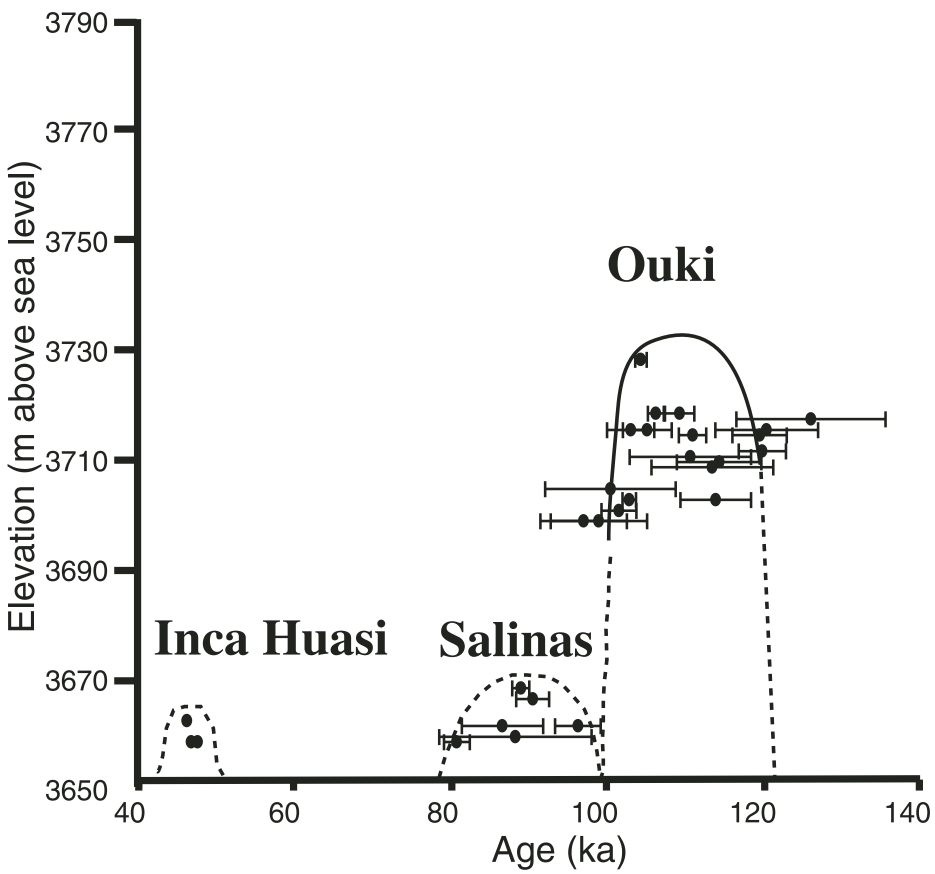

Figure 8. Reconstructed lake-level history between 140 and 40 ka. Error bars are not shown when smaller than the point representing the date.

Several sites mark continuous transgression of the Tauca lake cycle above 3670 m. An ooid crust (Fig. 3; site 18) at ~3670 m marks the shoreline of the transgressing lake and ^{14}C dates to 17,170 +570/–520 cal. yr B.P. A tufa in a small cave (Fig. 3; site 21; 3700 m) must have developed in very shallow water, as low light limited the depth of the photic zone in this cave. This sample U-Th dates to 16,880 ± 160 (^{14}C date of 17,030 +530/–490 cal. yr B.P.). A pisolitic tufa (Fig. 3; site 21), which likely formed in a high-energy (beach) environment U-Th dates to 17,950 ± 900. The very base of the Tauca Alloformation ^{14}C dates to 16,275 +490/–460 cal. yr B.P. where it rests directly on older basin fill at site 20 (Fig. 10A; 3725 m), and virtually identical ^{14}C results from Vinto (Fig. 3; 3725 m) put the base of the Tauca Alloformation at 16,370 +370/–490 cal. yr B.P. (Fig. 6).

Tauca Lake-Cycle Highstand

The highstand of the Tauca lake cycle is conspicuous throughout the basin as a series of 2–3 benches between 3765 and 3790 m (Fig. 5D). The large range of absolute elevation of these benches is an artifact of variable isostatic rebound (Bills et al., 1994). These benches, which we refer to as the Tauca shoreline complex, span ~10 m. The lower bench is the most prominent and is more encrusted by tufa. Tufas vary in height from one to at least five meters, and are mounded (Fig. 5E), which is a marked contrast to the conical cross sections of tufa in the Ouki Alloformation.

We established the duration of the Lake Tauca shoreline complex between 16.4 and 14.1 ka using nine ^{14}C and ten U-Th dates. These ^{14}C dates come from sediments exposed in natural cuts within 20 m of the shoreline complex from various locations around the basins. Shoreline deposits typically consist of cross-rippled, mollusk-rich sands locally mixed with diatomite resting on reddish hillslope colluvium and grading up into cemented beach rock. We captured the duration of the Tauca lake cycle by dating aquatic mollusks from the tops and bottoms of numerous Lake Tauca sequences exposed at various higher elevations around the basin. The ^{14}C ages (9 dates; Table 1) range from 16,370 +500/–460 (Fig. 3; site 3B; 3750 m) to 14,100 +1280/–290 cal. yr B.P. (Fig. 3; site 27; 3761 m). U-Th dates are from the lower highstand bench and fall between 16.3 and 13.3 ka; the dates with the lowest uncertainty are 15,540 ± 490 and 15,140 ± 300 (Table 3).

An age range of 16.4–14.1 ka for the highstand of Lake Tauca agrees with dates from the Tauca Alloformation at lower elevations. For example, the middle of the Tauca Alloformation at lower elevations ^{14}C dates to 15,800 +480/–1060 cal. yr B.P. (Fig. 3; Vinto; 3725 m), to 15,840 +480/–1070 (Fig. 3; site 20; 3725 m), to 15,440 +420/–1050 (Fig. 3; site 10; 3742 m), and to 15,230 +390/–1010 cal. yr B.P. (Fig. 3; site 9; 3710 m).

Tauca Lake-Cycle Regression

The regressional phase of the Tauca lake cycle is observed in several stratigraphic sections and tufa-encrusted benches at lower elevation. Many low-elevation benches, which probably represent short stillstands of the lake as it regressed, can be found throughout the basin. Tufa-encrusted benches at 3735 and 3740 m yield U-Th dates of 14,440 ± 340 (^{14}C: 13,940 +1325/–310 cal. yr B.P.) and 14,340 ± 270 (^{14}C: 14,290 +1170/–430 cal. yr B.P.). Deposits recording the drop in lake level are also widely exposed, usually consisting of mollusk-rich sands, often ripple cross-bedded or locally mixed with small tufa heads and carbonate-cemented gravel. There are too many dates from near the top of the Tauca Alloformation at elevations of 3745–3700 m to list individually. Eight ^{14}C dates in this category fall between 14,640 and 14,090 (Table 3), with large calibration errors of up to ±1350 yr. The only very low-elevation (Fig. 10K; 3661 m) sample that clearly dates the regression (^{14}C: 14,100 +1340/–580 cal. yr B.P.) comes from the top of carbonate-cemented beach gravel at Sajsi.

The U-Th and ^{14}C dates from the regressional phase overlap, within error, the latter part of the Lake Tauca highstand. This is due partly to the large calibration errors in ^{14}C during this time period, but it also shows that the regressional phase of the Tauca lake cycle was short, probably less than 500 yr.

The Coipasa Lake Cycle

We favor the presence of a separate, but minor, lake cycle between 13 and 11 ka, a lake cycle

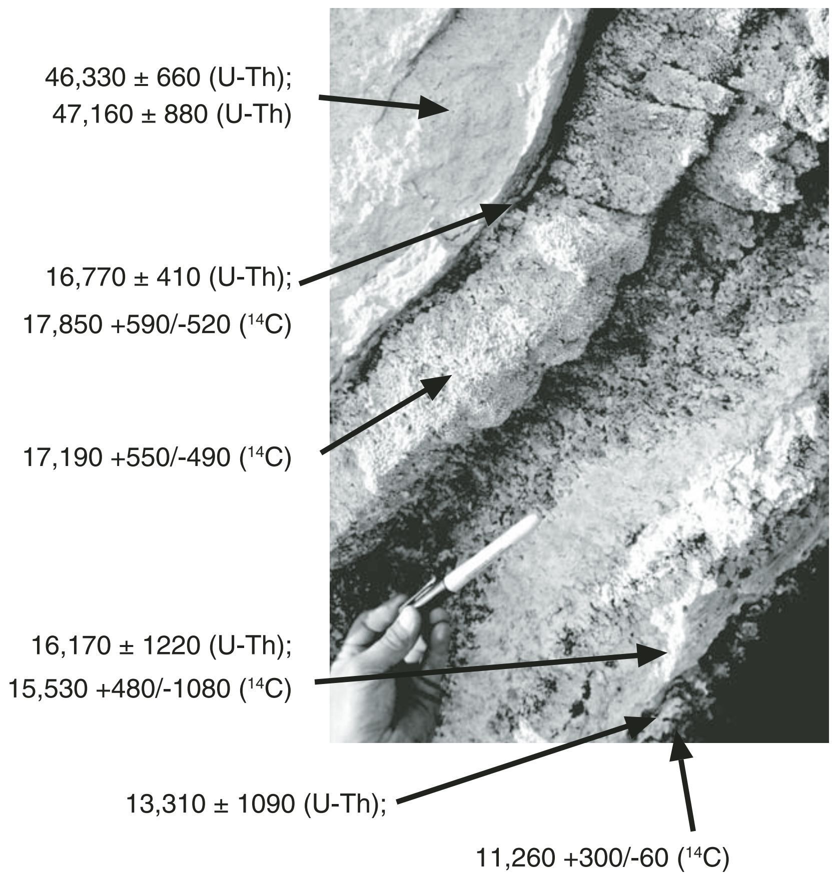

Figure 9. A layered carbonate from an island near the center of the Uyuni Basin (Fig. 3; site 19), with ^{14}C (cal. yr B.P.) and U-Th dates. Pen is 13.5 cm long and the top of the photograph is up. The many layers of Tauca tufa are often steeply inclined or mounded. Note the contrast between the clear disconformity between the Inca Huasi and Tauca layer and the ambiguous separation between the Tauca and Coipasa tufa layers.

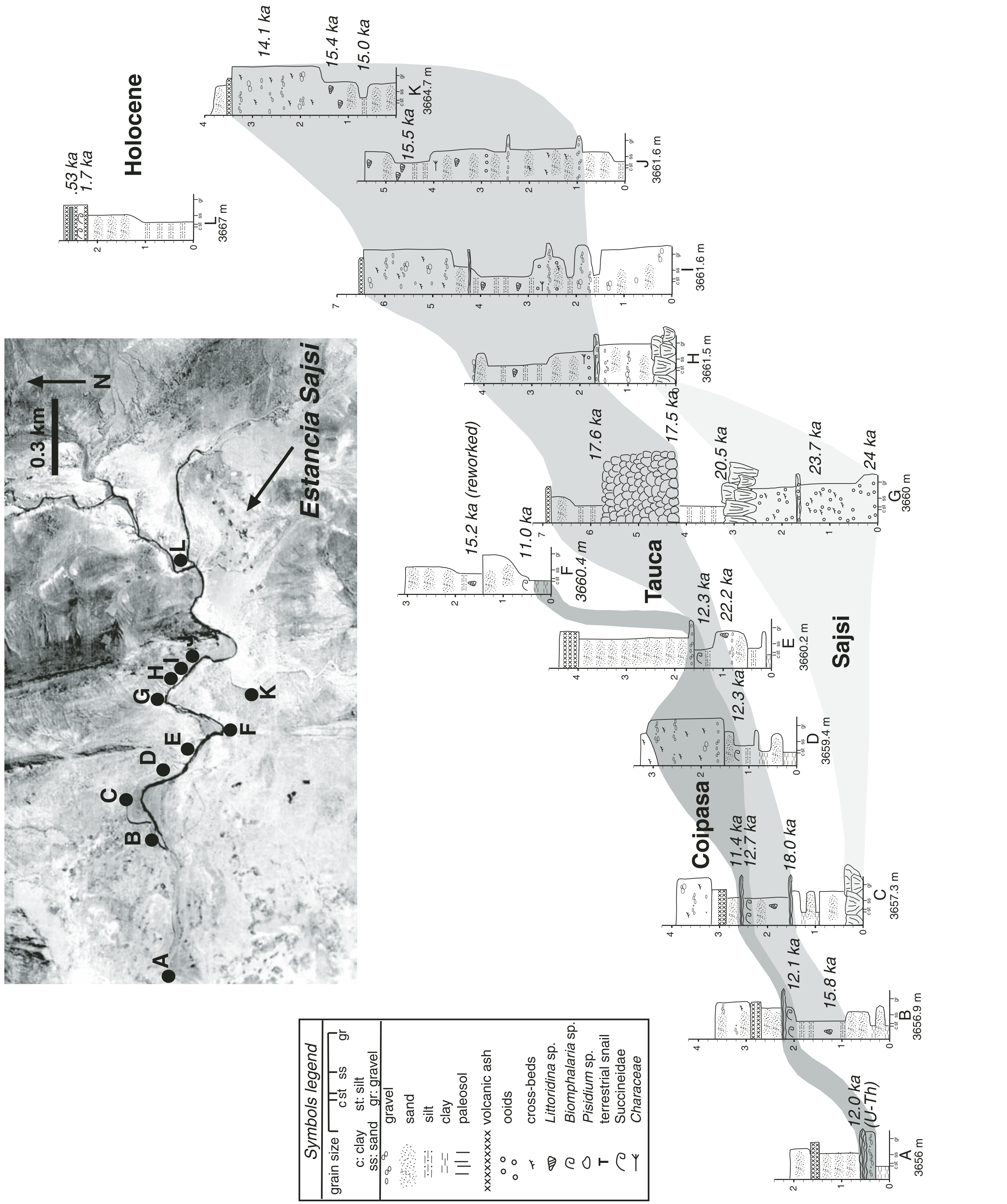
Symbols legend
grain size
c: clay
ss: sand
st: silt
gr: gravel
gravel
sand
silt
clay
paleosol
volcanic ash
ooids
cross-beds
Littoridina sp.
Biomphalaria sp.
Pisidium sp.
terrestrial snail
Succineidae
Characeae
0.3 km
N
Estancia Sajsi
A
B
C
D
E
F
G
H
I
J
K
L
Holocene
Tauca
Coipasa
Sajsi
12.0 ka (U-Th)
12.1 ka
15.8 ka
11.4 ka
12.7 ka
18.0 ka
12.3 ka
12.3 ka
22.2 ka
15.2 ka (reworked)
11.0 ka
17.6 ka
17.5 ka
20.5 ka
23.7 ka
24 ka
15.5 ka
14.1 ka
15.4 ka
15.0 ka
.53 ka
1.7 ka
3656 m
3656.9 m
3657.3 m
3659.4 m
3660.2 m
3660.4 m
3660 m
3661.5 m
3661.6 m
3661.6 m
3664.7 m
3667 m

Figure 10. The stratigraphy exposed at Sajsi (Fig. 3). Stratigraphic columns are keyed by letter to aerial view. Elevation for the base of each section is given. Radiocarbon dates are shown in thousands of calendar years (ka); one U-Th date is also given. The oldest lake deposits in this section belong to the Sajsi lake cycle. The transgression of Lake Sajsi is represented by oolitic sands. Shallow-water tufas with occasional ooids represent the deepest (<18 m deep) phase of Lake Sajsi, and the return of fluvial sand marks a dry period after 20.5 ka. The transgression of Lake Tauca is represented by a thin carbonate crust in the easternmost sections. The large delta formed when Lake Tauca was deep, and carbonate-cemented gravel dates the transgression of Lake Tauca. The carbonate crust attributed to the Coipasa lake cycle lies just above Succineidae-bearing sands (shallow water). Toward the salar, a beach gravel underlies the Coipasa tufa crust.

referred to originally by Sylvestre et al. (1999) as the Coipasa. Several dates, especially the carbonate-cemented gravel at Sajsi (Fig. 10K), indicate regression of Lake Tauca to very low elevations by ca. 14.1 ka, and a lake fluctuation (carbonate over beach gravel) of indeterminate depth is visible at Sajsi (Fig. 10A). The fresh, unweathered tufa crust (Fig. 5F) that is widespread below 3680 m is also attributed to the Coipasa lake cycle.

The best exposures of this lake cycle are at Sajsi (Figs. 3 and 11) in the eastern Uyuni Basin (Fig. 10). Above the Tauca Alloformation at low elevation (<3660 m) is sand and silt containing shells of Succineidae (^{14}C: 12,720 +240/–370; 12,320 +490/–540; 12,090 +640/–390, and 12,290 +500/–520 cal. yr B.P.), a semi-aquatic gastropod. This unit underlies a carbonate crust ^{14}C dating to 11,410 +530/–170 cal. yr B.P. that thickens basinward (Fig. 10A) and rests on an imbricated beach gravel above a sharp contact. In other places (Fig. 10B–D), the contact between the Tauca Alloformation and the Succineidae-rich unit is, however, gradual. A second Succineidae-bearing sand, less clearly lacustrine, dates by ^{14}C to 11,040 +120/–440 cal. yr B.P. In the same section (E), mollusks in alluvium (^{14}C: 15,390 +440/–1060 cal. yr B.P.) likely reworked from the Tauca Alloformation overlie the Succineidae-bearing sand.

A fresh white carbonate crust is commonly found at low elevations (<3680 m) around the Coipasa and Uyuni Basins. It coats older tufas and sediments and is distinctly less weathered. At lower elevations (<3660), this crust ^{14}C dates to between 13,480 +360/–390 and 12,260 +300/–60 cal. yr B.P. (Table 2). Although far less visible at higher elevation, chronologic constraints suggest that this same crust extends to ~3700 m where we obtained U-Th ages of 12,260 ± 250 and 12,260 ± 330 (Fig. 3; site 15), and 12,260 ± 180 (Fig. 3; site 25; ^{14}C dates: 11,040 +120/–400 and 11,260 +430/–80 cal. yr B.P.).

We attribute this young carbonate crust and the youngest sedimentary sequence at Sajsi to the Coipasa lake cycle. The Succineidae-rich unit at Sajsi with dates ranging between 12.7 and 12.3 ka is indicative of very shallow water at Sajsi, as this species of snail is semi-aquatic, living on vegetation just above the water (Pigati et al., 2004). The extensive nature of this Succineidae-rich unit is likely indicative of formation along the transgressing lake shore or deltaic wetland. The tufa overlying this unit formed in slightly deeper water, and we tentatively tie this crust (with a U-Th date of 11,970 ± 1290) to similar looking crusts and tufa coatings at higher elevations. We tentatively date the end of the Coipasa lake cycle to 11,040 +120/–440 cal. yr B.P., the date of the second Succineidae-bearing sand at Sajsi.

Unlike the break between Inca Huasi and Tauca tufas, the difference between Tauca and Coipasa layers within tufa encrustation is, however, not always clear (Fig. 9). Thus, we observe a lake oscillation at Sajsi and tie this oscillation to higher elevation tufas on the basis of tufa texture and chronological constraints. We stress that our Coipasa lake cycle chronology is very tentative, and one ^{14}C date of 13,180 +600/–160 cal. yr B.P. (Fig. 3; site 3A) at high elevations is problematic.

Holocene Deposits

Thin alluvial sands and eolian silts rest directly upon the Tauca Alloformation above ~3700 m and on Coipasa-age deposits at lower elevation. Where not eroded, the Hanco Geosol overprints these deposits. The Hanco Geosol consists of a weakly developed, tan (10 YR 4/4) cambic or argillic horizon underlain by a weak stage II carbonate horizon (sensu Gile et al., 1966).

Radiocarbon dates from nonlacustrine deposits near the Salars de Uyuni and Coipasa provide some important constraints on Holocene lake-level history. Archaeological charcoal from a rocky shelter just meters above Salar de Coipasa (Fig. 3; site 7; 3660 m) yielded two ^{14}C dates of 6238 +60/–237 and 6372 +70/–87 cal. yr B.P. We also obtained a number of ^{14}C dates from postlacustrine marsh deposits across the basin. These include 8,404 +134/–65 (Fig. 3; site 9) at 3718 m and 2330 +40/–180 cal. yr B.P. at 3705 m (Fig. 3; site 8) from terrestrial snails from fluvial deposits, and 1710 +154/–139 cal. yr B.P. from semi-aquatic snails (also 530 +91/–21 cal. yr B.P.) (Fig. 10L). Thick paleowetland deposits located above the Lake Tauca shoreline at site 26 (Fig. 3) ^{14}C date between 10,610 +130/–190 and 3180 +170/–180 cal. yr B.P.

These deposits provide some constraints on lake elevation and recent climate in the basin. For example, no lacustrine deposits overlie the archaeological strata (Fig. 3; site 7) found at Salar de Coipasa, so the lake must have remained below 3660 m (i.e., no more than 8 m deep) since ca. 6.4 ka. The thick and extensive paleowetland deposits (Fig. 3; site 26) dating between 10.6 and 3.2 ka are consistent with another record of regional wetlands expansion in the basins (Servant and Servant-Vildary, 2003), but the climatic significance of Altiplano wetland development at a time of very low lake levels remains unclear.

SYNTHESIS OF THE LAKE RECORD

Paleoshorelines: Climate or Basin Geometry?

Climate impacts lake levels in closed basins by altering the hydrologic balance between runoff, precipitation, and evaporation, but basin geometry also influences lake levels by altering the surface area to volume ratio. In large lake systems elsewhere (e.g., Bonneville, Lahontan, and Lisan), well-developed shorelines correspond to periods when lake level was stabilized as a result of spilling over into an arid receiving basin (Curry and Oviatt, 1985; Benson and Paillet, 1989; Benson et al., 1990; Bartov et al., 2002). In the Poopo, Coipasa, and Uyuni Basins, a major strandline occurs at ~3700 m, the elevation of the Laka sill, and may correspond to the Coipasa lake-cycle highstand. Likewise, the two main benches in the Tauca shoreline complex are similar in elevation to spillways into two nearby subbasins, Ascotan and Laguani (Fig. 2). These subbasins, however, have surface areas that constitute <1% of the ~60,000 km^2 area of the Lake Tauca (Blodgett et al., 1997) and, by themselves, are unlikely to have impeded the transgression of Lake Tauca long enough to create significant strandlines. The well-developed benches associated with the Lake Tauca highstand must have been produced by a lake already close to hydrologic steady state, but that was further stabilized by the increase in surface area provided by overflow into these small subbasins.

Lake Tauca regressional benches at ~3740 m are minor compared to the Tauca shoreline complex. The Empexa Basin is more than twice the size of Ascotan and connects to the Uyuni Basin at ~3735 m, an elevation at which Lake Tauca had a surface area of less than 30,000 km^2 (Blodgett et al., 1997). Evidently, Lake Tauca rose and fell relatively rapidly in response to climate change, leaving very little evidence of a standstill at the ~3735 m divide between Uyuni and Empexa. We do note, however, that the Lake Ouki highstand is at roughly 3735 m.

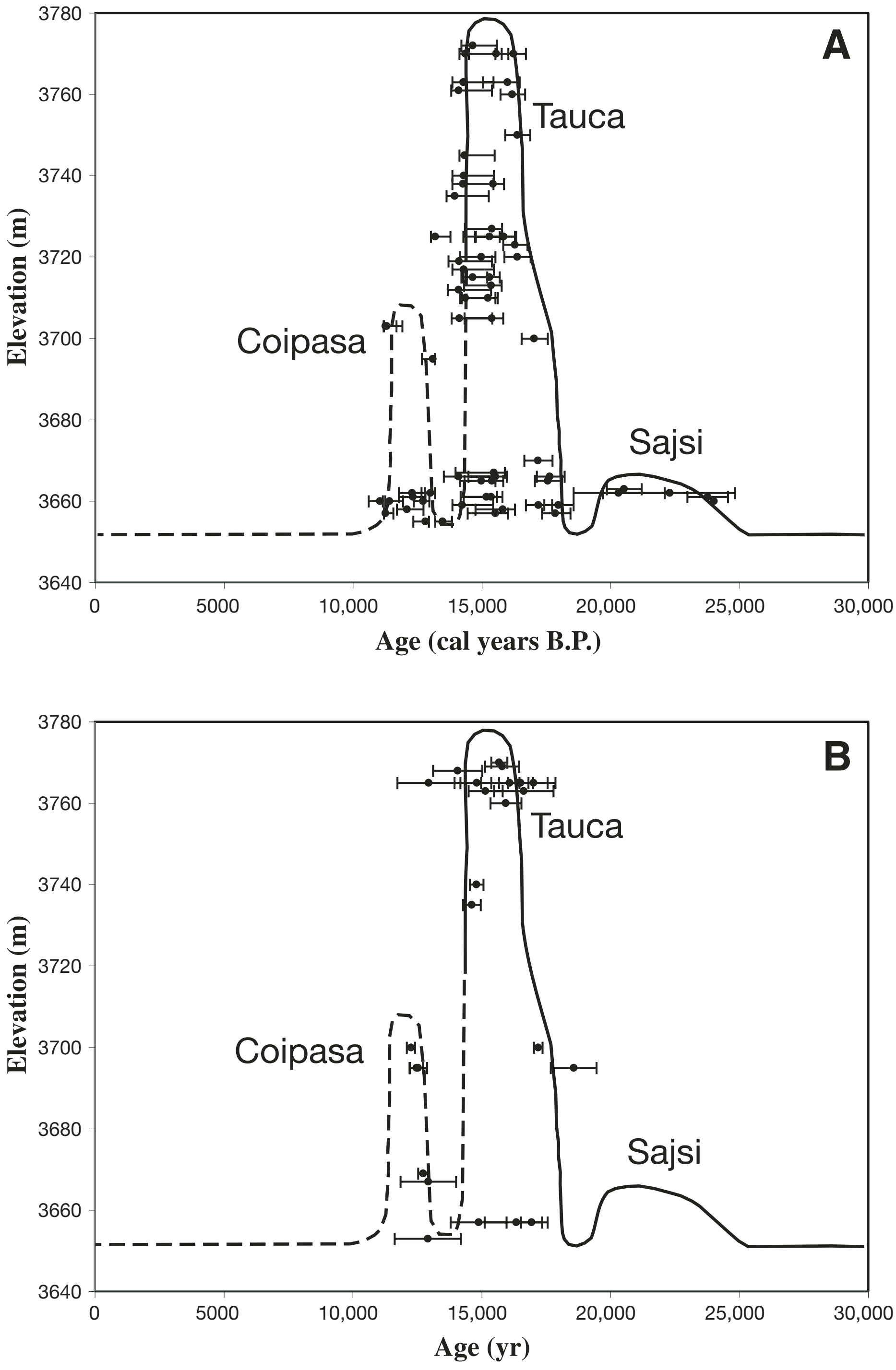

Figure 11. Proposed lake-level curve for the Sajsi, Tauca, and Coipasa lake cycles with ^{14}C (A) and U-Th (B) dates.

Comparison to Other Records

Our study provides the first geochronologic evidence for three previously unrecognized lake cycles in shoreline stratigraphy of the basin: the Ouki (ca. 120–98 ka), Inca Huasi (ca. 47 ka), and Sajsi lake cycles (24–20.5 ka). We also find that lake deposits previously attributed to the "Minchin" probably belong to several different lake cycles, in particular the Ouki lake cycle. We favor dropping the term "Minchin," because we now know it probably subsumes several lake cycles.

Our chronology is consistent with previous reconstructions of lake history based on shoreline and short-core evidence over the last 18 k.y. (the Tauca and Coipasa lake cycles). Sylvestre et al. (1999) reported 26 ^{14}C dates from shoreline deposits and short-sediment cores for the Tauca lake cycle, all falling between 18.5 and 14.0 ka, which match our estimate of 18.1–14.1 ka for the duration of the Tauca lake cycle. Diatom evidence from the same study points to a Lake Tauca highstand shortly after 16.4 ka, overlapping the ^{14}C dates of 16,520 +500/–1700 cal. yr B.P. obtained by Bills et al. (1994).[2] Six U-Th dates from Sylvestre et al. (1999) range between 18.8 and 10.9 ka for Lake Tauca, although only one of these dates (15,070) displays low initial Th, comparable to the U-Th dates that we report. Nine ^{14}C dates for the Coipasa lake cycle range between 13.4 and 12.5 ka (Sylvestre et al., 1999), partially overlapping our age estimate (13–11 ka) for this lake cycle.

Ideally, shoreline deposits and long-sediment cores provide complementary paleohydrologic information about the history of lake-level changes. Sediment core records are generally more continuous and should capture both major and minor lake events. However, they provide only indirect, proxy evidence of lake depth. By contrast, shoreline stratigraphy provides direct physical evidence of lake depth and allows for replication of dates.

Two long cores have been studied from Uyuni Basin (Fornari et al., 2001; Risacher and Fritz, 2000; Baker et al., 2001a; Fritz et al., 2004). The cores have similar stratigraphy, consisting

[2]An extensive discussion of the potential causes of the discrepancy between the two ^{14}C dates of Bills et al. (1994) and the chronology of Sylvestre et al. (1999) is found in Sylvestre et al. (1999). We find, however, that the dates of Bills et al. (1994) are consistent with our chronology and that of Sylvestre et al. (1999) when calibrated using CALIB 3.4. The original calibrated age of these dates (^{14}C age of 13,790 ± 70) is 16.8 ka, using the calibration of Stuiver and Reimer (1993). Our calibrated age for these dates, using CALIB 3.4, is 16,520 +500/–1700 cal. yr B.P.

of alternating mud and salt layers. Mud units are attributed to paleolakes and salt units to dry intervals when the Uyuni Basin was occupied by a salar. Eleven mud (lake) and 12 salt (desiccation) units are present in both cores; we refer to the mud units as L1, L2, L3, etc., from youngest to oldest, after Risacher and Fritz (2000). The geochronology for these cores differs dramatically, making it difficult to compare all but the youngest layers to our results from shorelines. Until these geochronologic differences are resolved, we will confine our discussion to the post–30 ka chronology only.

The first two mud intervals, L1 and L2, in the Uyuni sediment cores correlate with the Coipasa and Tauca lake cycles, respectively. In the core studied by Fornari et al. (2001), ^{14}C dates from organic matter in L1 place it at 12,990 +160/–280 cal. yr B.P., similar in age to our Coipasa lake cycle. The same interval was not recovered from the core studied by Baker et al. (2001a) and Fritz et al. (2004). The base of L2 corresponds to a sharp drop in halite content in one core, which Fornari et al. (2001) ^{14}C dated to 16,100 +2900/–2300 cal. yr B.P. This matches our estimate for the start of the Tauca deep-lake cycle. In contrast, Baker et al. (2001a) defined the extent of L2 using elevated natural gamma radiation values, an interval that ^{14}C dates between 14,900 and 26,100 cal yr B.P. (Fig. 12B), which encompasses both the shallow Sajsi and much deeper Tauca lake cycles. Diatom proxies for depth and salinity are unavailable over the last ~20 k.y. from the core studied by Fritz et al. (2004) and Baker et al. (2001a), but such evidence from diatoms is available from a short core in the Coipasa Basin and shows that the lake was very shallow between 22 and 18 ka (Sylvestre, 2002).

Lake Titicaca, together with the Poopo, Coipasa, and Uyuni Basins, spans more than seven degrees of longitude (Fig. 1). Lake Titicaca currently loses ~10% of its annual water budget by outflow to the Rio Desaguadero (Roche et al., 1992). The relative contribution of paleooutflow from Lake Titicaca to the southern basins, however, is disputed (Coudrain et al., 2002; Grove et al., 2003). Whatever the relative contribution, several proxies indicate that Lake Titicaca was overflowing from ca. 26 to after 15 ka (Baker et al., 2001b) (Fig. 12C), which is consistent with our results.

Glacial advances and lake-level increases in the Poopo, Uyuni, and Coipasa Basins are generally synchronous. Lake sediments from the Tauca deep-lake cycle are interbedded with glacial outwash, and this major regional glacial advance is considered synchronous with the growth of Lake Tauca (Clapperton et al., 1997; Clayton and Clapperton, 1997). The extent of paleolakes on the Altiplano over the last 24 k.y. may be recorded by variations in anion concentration on Mount Sajama (Fig. 12D; Thompson et al., 1998). Ice cores from Mount Sajama are not directly dated between 10 and 24 ka, but decreases in Cl^- concentration broadly correspond to lake expansions. Changes in accumulation rates on Sajama, however, do not always covary with anion concentrations (Fig. 12D; Thompson et al., 1998).

DISCUSSION

Comparisons between our new chronology and global climate allow us to reconsider the

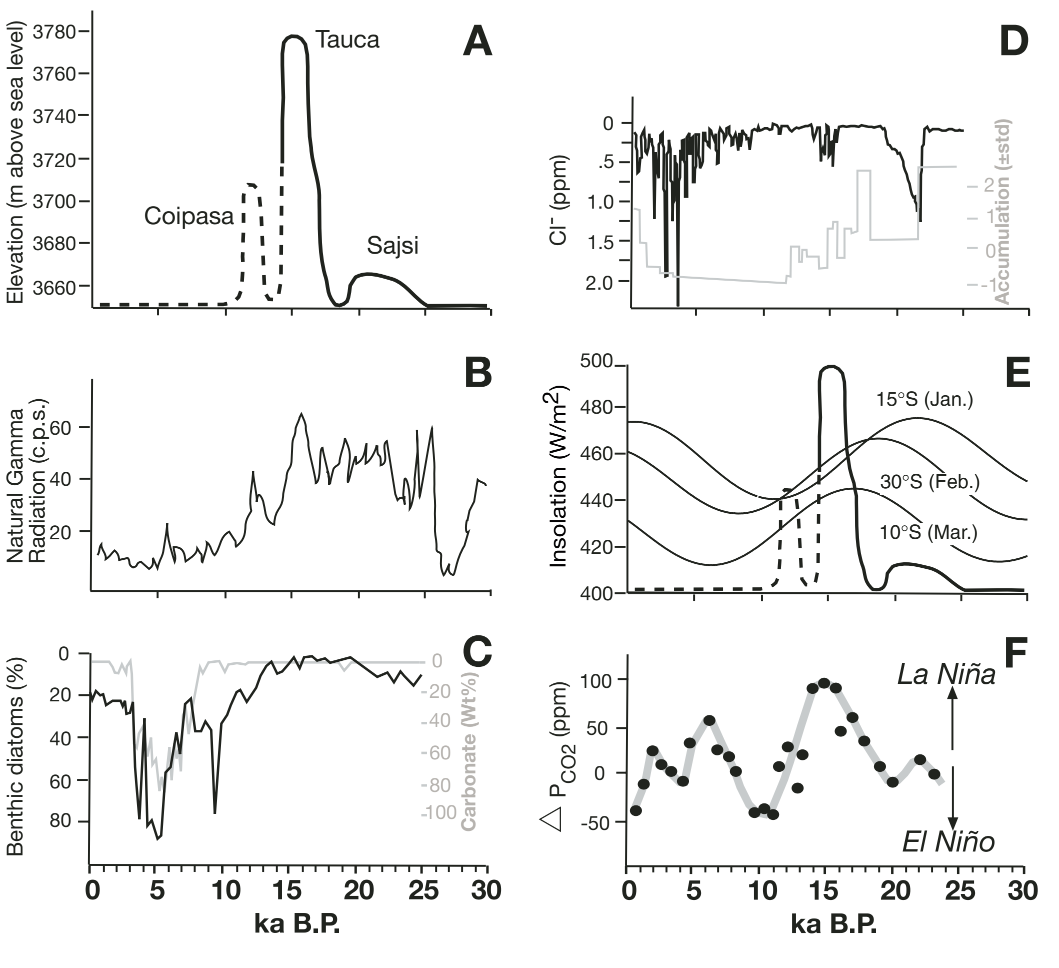

Figure 12. Comparison of paleohydrologic and climate proxies over the last 30 k.y. (A) reconstructed lake-level curve (this study); (B) natural gamma radiation (counts per second) from an Uyuni core (Baker et al., 2001a); (C) percent benthic diatoms (black) and weight percent carbonate (gray) from core 1PC in Lake Titicaca (Baker et al., 2001b); (D) chlorine concentration (black) and accumulation rate (gray) of ice on Mount Sajama (Thompson et al., 1998); (E) reconstructed lake-level curve (this study) and insolation curves (Baker et al., 2001a; Wang et al., 2004; Cruz et al., 2005) proposed as drivers of a southward shift in the Intertropical Convergence Zone (ITCZ); (F) change in P_{CO_2} in the western equatorial Pacific inferred from boron isotope analyses of planktonic foraminifera, in which increased P_{CO_2} is associated with stronger upwelling and La Niña–like conditions (Palmer and Pearson, 2003).

cause of Altiplano lake fluctuations. Four potential causes have been proposed: (1) changes in seasonality, especially local summer insolation (Baker et al., 2001a, 2001b; Rowe et al., 2002; Fritz et al., 2004); (2) changes in global temperature (Blodgett et al., 1997; Garreaud et al., 2003); (3) changes in aridity over the Amazon Basin (Mourguiart and Ledru, 2003); and (4) changes in SST gradients (Betancourt et al., 2000; Baker et al., 2001a, 2001b; Garreaud et al., 2003).

Our new chronology strongly argues against simple forcing of the South American summer monsoon resulting from changing local January insolation. In theory, heating over the high Altiplano might intensify the South American summer monsoon directly by increasing the thermal contrast between the Altiplano and the Atlantic and/or less directly by southward displacement of the convection associated with the ITCZ. We rule out local January insolation as the primary driver of lake cycles, because both deep lakes occurred during periods of low to moderate local summer insolation. Lake Tauca reached a maximum between 16.4 and 14.1 ka, ~5 k.y. after the insolation peak at ca. 20 ka (Fig. 12E), and Lake Ouki spanned the most profound minimum (105–100 ka) in January insolation in the last 200 k.y. March isolation, with peaks at ca. 17 ka, 40 ka, 66 ka, 88 ka, and 111 ka is, however, a relatively good fit with our lake record (Fig. 13B). The position of the ITCZ through time may eventually be linked to precessional cycling, but such conclusions await additional geographically distributed records from the Amazon Basin. The relative amplitude and timing of the Sajsi, Tauca, and Coipasa lake cycles cannot, however, be explained by any solar insolation curve.

At other locations in the Amazon Basin (e.g., Cruz et al., 2005; Wang et al., 2004), local insolation at various seasons may drive local convection and/or the position of the ITCZ, although no single season or latitude of precessional cycling seems to fit the evidence. Two long (more than 100 k.y.) speleothem-based records of climate from the Amazon Basin point to precession-driven changes in solar insolation as a principal driver of shifts in the north-south position of the ITCZ (Wang et al., 2004; Cruz et al., 2005). There is a pronounced ~23 k.y. cycle in the record of Cruz et al. (2005), and February insolation at 30°S is in-phase with wet and dry cycles over this site (Wang et al., 2004). In contrast, austral autumn insolation at 10°S is linked to wet periods in northeast Brazil (Wang et al., 2004). The timing of the Ouki lake cycle is consistent with the record from Botuverá Cave (30°S), but the Tauca lake cycle is synchronous with increased moisture in northeast Brazil (10°S).

Pole-to-equator temperature gradients may also play a significant role in the position of the ITCZ; such gradients may be tied to Northern Hemisphere ice volumes (e.g., Chiang et al., 2003), Atlantic SST gradients (e.g., Chang et al., 1997), and even ENSO (e.g., Haug et al., 2001). Our chronology tentatively places the Coipasa lake cycle between 13 and 11 ka, roughly coincident with the Younger Dryas.

All else being equal, cold temperatures should favor lake expansion by decreasing evaporation rates, and warm global temperatures should strengthen the South American summer monsoon by intensifying the thermal contrast between the Altiplano and the Atlantic Ocean. Deep-lake cycles on the southern Altiplano, however, occur during periods of moderate global temperature, so neither decreased evaporation nor warming of the Altiplano appears to dominate lake expansion. A temperature decrease of 10 °C is required to reduce evaporation rates sufficiently to produce a lake the size of Lake Tauca (Blodgett et al., 1997). A temperature decrease of this magnitude is inconsistent with the timing of lakes Ouki and Tauca, leaving increased precipitation as the main alternative for producing deep lakes. Global temperature changes may, however, play an indirect role in lake expansion by modulating moisture over the Amazon Basin.

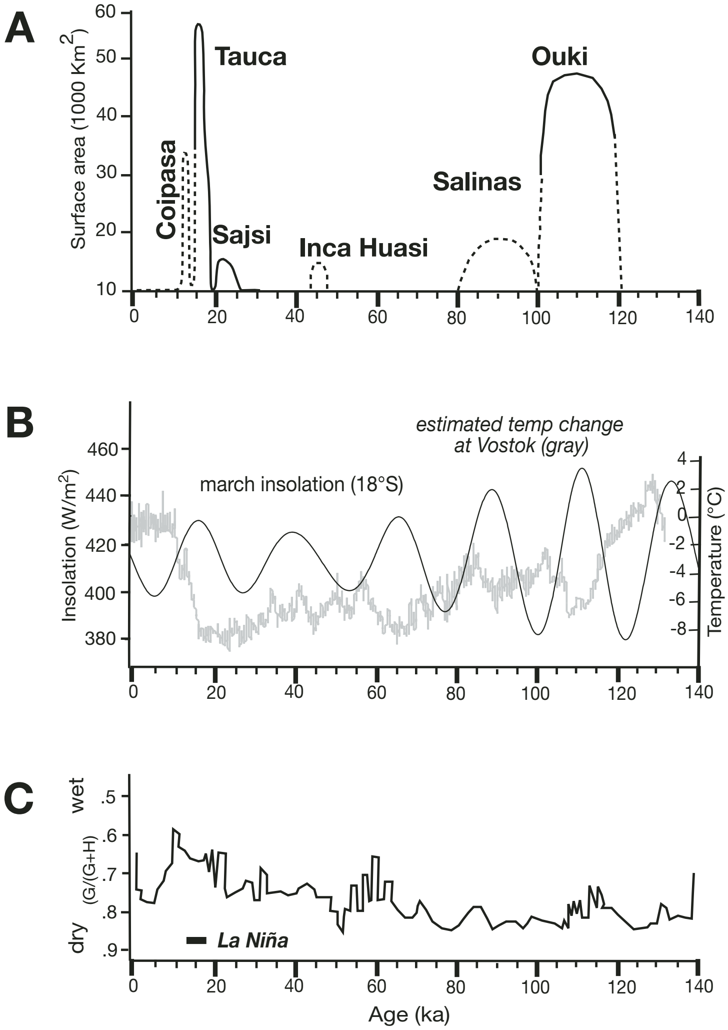

Figure 13. Comparison of regional paleoclimate proxies from South America: (A) reconstructed lake history from shoreline deposits. Surface area is calculated using elevational data from Blodgett et al. (1997). (B) March insolation at 18°S (black) (Laskar, 1990) and estimated temperature change at Vostok (gray) (Petit et al., 1999). (C) Iron oxide composition (goethite/[goethite + hematite]) of sediments derived from the Amazon (Harris and Mix, 1999). La Niña–like conditions (Palmer and Pearson, 2003) are noted between 18 and 13 ka.

Temperature may also directly impact minor lake fluctuations, such as the Sajsi event.

Overall, aridity on the Bolivian Altiplano and aridity in the Amazon appear to be linked, with cold glacial intervals being relatively dry (Fig. 13C). Many studies (e.g., Harris and Mix, 1999; Behling, 2002; Garreaud et al., 2003) suggest that the Amazon Basin was generally dry during glacial periods. Only the minor Inca Huasi and Sajsi lake cycles occur between 18.1 and 80 ka, suggesting that glacial epochs were also relatively dry on the Altiplano. Today, the water content of air over the Gran Chaco region of Argentina influences rainfall on the southern Altiplano (Vuille and Keimig, 2004), and under more arid conditions, the Amazon may exert greater control over precipitation on the Altiplano. Records from the Bolivian highlands have recently been used to argue that the Amazon Basin was both wet (Baker et al., 2001b) and dry (Mourguiart and Ledru, 2003) during the Last Glacial Maximum (LGM). The presence of shallow lake Sajsi during this period (ca. 22 ka) argues for a LGM that was relatively dry.

We suggest that important ancient links may exist between central Andean moisture and Pacific SST gradients, the primary driver of modern interannual variability in the region. The timing of the Tauca deep-lake cycle coincides with evidence for intense upwelling in the central Pacific between 18 and 13 ka (Palmer and Pearson, 2003) (Fig. 12E). Such upwelling is indicative of strong La Niña–like conditions, which today result in wet years on the Altiplano. The modern relationship between ENSO and the strength of trade winds and/or the position of the ITCZ is also evident throughout the Holocene in the Cariaco Basin, northern Venezuela (Haug, et al., 2001). Climate records of Pacific SST variations during the Ouki deep-lake cycle (ca. 120–98 ka) are not well developed. Thus, the possibility that the Ouki cycle, like the Tauca deep-lake cycle, is linked to persistent La Niña–like conditions awaits analysis of further records of SST gradients from the Pacific.

CONCLUSIONS

This new shoreline chronology provides an ~120 k.y. record of moisture balance on the Altiplano. Over this time, the Poopo, Coipasa, and Uyuni basins were twice occupied by lakes with surface areas in excess of 30,000 km^2. The Ouki deep-lake cycle reached its maximum elevation (3740 m) between 110 and 100 ka, but may have commenced as early as 120 ka and ended as late as 98 ka. Much smaller lakes were present in the Uyuni Basin between ca. 95 and 80 ka (Salinas lake cycle), around 46 ka (Inca Huasi lake cycle), and between 24 and 20.5 ka (the Sajsi lake cycle). Lake Sajsi receded to <5 m depth before the Tauca deep-lake cycle transgressed at 18.1 ka. Lake Tauca reached its highstand (~60,000 km^2) between 16.4 and 14.1 ka. We tentatively date the Coipasa lake cycle between 13 and 11 ka.

The timing of the Altiplano's older deep-lake cycle at 120–98 ka, instead of the 28–50 ka indicated by contaminated "Minchin" ^{14}C dates, calls for a fundamental shift in explanations of climate change in the region. Shoreline evidence no longer favors the presence of a very large lake during the 20–80 ka cold period, and both deep-lake cycles on the Altiplano are out-of-phase with local summer insolation.

Lake cycles on the Bolivian Altiplano are influenced by both moisture content in the Amazon Basin and enhanced transport of that moisture onto the Altiplano by a favorable position of the ITCZ and/or by La Niña–like Pacific SST gradients. Glacial epochs (including the LGM) were relatively dry on the Altiplano, with only the minor Inca Huasi and Sajsi lake cycles between 80 and 18.1 ka. Climate conditions in the Amazon Basin over this interval remain disputed, but this new record implies that the distal end of the Amazon moisture source was dry throughout this time. Expansion of Lake Tauca coincides with La Niña–like conditions in the Pacific, suggesting ancient links between central Andean moisture and Pacific SST gradients, the primary driver of modern interannual precipitation variability on the Bolivian Altiplano.

ACKNOWLEDGMENTS

We thank Sohrab Tawakholi and Servicio Nacional de Geologia y Mineria (SERGEOMIN) for field logistical support; the University of Arizona Accelerator Facility for ^{14}C dates; N. English and J. Mendez for field assistance; W. Pratt for mollusk identifications; G. Gehrels, W. Beck, and I. Winograd for reference materials and samples; C. Isachsen, J. Wagner, G. Manhes, Z. Naiman, and T. Fischer for advice in the laboratory; and J. Betancourt, B. Bills, and K. Adams for extensive discussions. This work was supported by the National Science Foundation (NSF) grant EAR-0207850 to Quade and Patchett, and by grants from the Geological Society of America, the Arizona Geological Survey, Chevron, and University of Arizona Accelerator Facility. We also appreciate the comments of Florence Sylvestre and Donald Rodbell.

REFERENCES CITED

Aceituno, P., 1988, On the functioning of the Southern Oscillation in the South America sector, Part I: Surface climate: Monthly Weather Review, v. 116, p. 505–524, doi: 10.1175/1520–0442(1989)002<0341:OTFTS>2.0.CO;2.

Aceituno, P., and Montecinos, A., 1993, Circulation anomalies associated with dry and wet periods in the South American Altiplano, *in* Proceedings of the Fourth International Conference on Southern Hemisphere Meteorology: Hobart, Australia, American Meteorological Society, p. 330–331.

Ahlfeld, F., and Branisa, L., 1960, Geologia de Bolivia: La Paz, Editorial Don Bosco, 245 p.

Argollo, J., and Mourguiart, P., 2000, Late Quaternary climate history of the Bolivian Altiplano: Quaternary International, v. 72, p. 37–51, doi: 10.1016/S1040-6182(00)00019-7.

Baker, P.A., Rigsby, C.A., Seltzer, G.O., Fritz, S.C., Lowenstein, T.K., Bacher, N.P., and Veliz, C., 2001a, Tropical climate change at millennial and orbital timescales on the Bolivian Altiplano: Nature, v. 409, p. 698–701, doi: 10.1038/35055524.

Baker, P.A., Seltzer, G.O., Fritz, S.C., Dunbar, R.B., Grove, M.J., Tapia, P.M., Cross, S.L., Rowe, H.D., and Broda, J.P., 2001b, The history of South American tropical precipitation for the past 25,000 years: Science, v. 291, p. 640–643, doi: 10.1126/science.291.5504.640.

Bard, E., 1998, Geochemical and geophysical implications of the radiocarbon calibration: Geochimica et Cosmochimica Acta, v. 62, p. 2025–2038, doi: 10.1016150016-7037(98)00130-6.

Bartov, Y., Stein, M., Enzel, Y., Agnon, A., and Reches, Z., 2002, Lake levels and sequence stratigraphy of Lake Lisan, the Pleistocene precursor of the Dead Sea: Quaternary Research, v. 57, p. 9–21, doi: 10.1006/qres.2001.2284.

Behling, H., 2002, South and southeast Brazilian grasslands during Late Quaternary times: A synthesis: Palaeogeography, Palaeoclimatology, Palaeoecology, v. 177, p. 19–27, doi: 10.1016/S0031-0182(01)00349-2.

Benson, L.V., and Paillet, F.L., 1989, The use of total lake-surface area as an indicator of climatic change: Examples from the Lahontan Basin: Quaternary Research, v. 32, p. 262–275, doi: 10.1016/0033-5894(89)90093-8.

Benson, L.V., Currey, D.R., Dorn, R.I., Lajoie, K.R., Oviatt, C.G., Robinson, S.W., Smith, G.I., and Stine, S., 1990, Chronology of expansion and contraction of four Great Basin lake systems during the past 35,000 years: Palaeogeography, Palaeoclimatology, Palaeoecology, v. 78, p. 241–286, doi: 10.1016/0031-0182(90)90217-U.

Betancourt, J.L., Latorre, C., Rech, J.A., Rylander, K.A., and Quade, J., 2000, A 22,000-year record of monsoonal precipitation from northern Chile's Atacama Desert: Science, v. 289, p. 1542–1546, doi: 10.1126/science.289.5484.1542.

Bills, B.J., deSilva, S.L., Currey, D.R., Emenger, R.S., Lillquist, K.D., Donnellan, A., and Worden, B., 1994, Hydro-isostatic deflection and tectonic tilting in the central Andes: Initial results of a GPS survey of Lake Minchin shorelines: Geophysical Research Letters, v. 21, p. 293–296, doi: 10.1029/93GL03544.

Bird, M.I., Ayliffe, L.K., Fifield, L.K., Turney, C.S.M., Cresswell, R.G., Barrows, T.T., and David, B., 1999, Radiocarbon dating of "old" charcoal using a wet oxidation, stepped-combustion procedure: Radiocarbon, v. 41, p. 127–140.

Blodgett, T.A., Lenters, J.D., and Isacks, B.L., 1997, Constraints on the origin of paleolake expansions in the central Andes: Earth Interactions 1: http://ams.allenpress.com/amsonline/?request-get-archive&issn-1087-3562 (last accessed march, 2001).

Brennan, R., and Quade, J., 1997, Reliable late-Pleistocene stratigraphic ages and shorter groundwater travel times from ^{14}C in fossil snails from the southern Great Basin: Quaternary Research, v. 47, p. 329–336, doi: 10.1006/qres.1997.1895.

Cane, M.A., 1998, A role for the tropical Pacific?: Science, v. 282, p. 59–61.

Chang, P., Ji, L., and Li, H., 1997, A decadal climate variation in the tropical Atlantic Ocean from thermodynamic air-sea interactions: Nature, v. 385, p. 516–518, doi: 10.1038/385516a0.

Chiang, J.C.H., Biasutti, M., and Battisti, D.S., 2003, Sensitivity of the Atlantic ITCZ to Last Glacial Maximum boundary conditions: Paleoceanography, v. 18, p. 1094, doi: 10.1029/2003PA000916.

Clapperton, C.M., Clayton, J.D., Benn, D.I., Marden, C.J., and Argollo, J., 1997, Late Quaternary glacier advances and palaeolake highstands in the Bolivian Altiplano: Quaternary International, v. 38-39, p. 49–59, doi: 10.1016/S1040-6182(96)00020-1.

Clayton, J.D., and Clapperton, C.M., 1997, Broad synchrony of a late-glacial glacier advance and the highstand of palaeolake Tauca in the Bolivian Altiplano:

Journal of Quaternary Science, v. 12, p. 169–182, doi: 10.1002/(SICI)1099-1417(199705/06)12:3<169::AID-JQS304>3.0.CO;2-S.

Colinvaux, P.A., De Oliveira, P.E., and Bush, M.B., 2000, Amazonian and neotropical plant communities on glacial time-scales: The failure of the aridity and refuge hypotheses: Quaternary Science Reviews, v. 19, p. 141–169, doi: 10.1016/S0277-3791(99)00059-1.

Coudrain, A., Loubet, M., Condom, T., Talbi, A., Ribstein, P., Pouyaud, B., Quintanilla, J., Dieulin, C., and Dupre, B., 2002, Données isotopiques (^{87}Sr/^{86}Sr) et changements hydrologiques depuis 15000 ans sur l'Altiplano Andin: Hydrological Sciences, v. 47, p. 293–306.

Cruz, F.W., Burns, S.J., Karmann, I., Sharp, W., Vuille, M., Cardoso, A.O., Ferrari, J.A., Silva Dias, P.L., and Viana, O., 2005, Insolation-driven changes in atmospheric circulation over the past 116,000 years in subtropical Brazil: Nature, v. 434, p. 63–66, doi: 10.1038/nature03365.

Curry, D.R., and Oviatt, C.G., 1985, Duration, average rates, and probable causes of Lake Bonneville expansions, still-stands, and contractions during the last deep-lake cycle, 32,000 to 10,000 years ago, *in* Kay, P.A., and Diaz, H.F., eds., Problems of and prospects for predicting Great Salt Lake levels: Salt Lake City, University of Utah Center for Public Affairs and Administration, p. 9–24.

Fornari, M., Risacher, F., and Feruad, G., 2001, Dating paleolakes in the central Altiplano of Bolivia: Palaeogeography, Palaeoclimatology, Palaeoecology, v. 172, p. 269–282, doi: 10.1016/S0031-0182(01)00301-7.

Fritz, S.C., Baker, P.A., Lowenstein, T.K., Seltzer, G.O., Rigsby, C.A., Dwyer, G.S., Tapia, P.M., Arnold, K.K., Ku, T.-L., and Luo, S., 2004, Hydrologic variation during the last 170,000 years in the Southern Hemisphere tropics of South America: Quaternary Research, v. 61, p. 95–104, doi: 10.1016/j.yqres.2003.08.007.

Garreaud, R.D., and Aceituno, P., 2001, Interannual rainfall variability over the South American Altiplano: Monthly Weather Review, v. 125, p. 3157–3171.

Garreaud, R.D., Vuille, M., and Clement, A.C., 2003, The climate of the Altiplano: Observed current conditions and mechanisms of past changes: Palaeogeography, Palaeoclimatology, Palaeoecology, v. 194, p. 5–22, doi: 10.1016/S0031-0182(03)00269-4.

Geyh, M., Grosjean, M., Núñez, L., and Schotterer, U., 1999, Radiocarbon reservoir effect and the timing of the late-glacial/early Holocene humid phase in the Atacama Desert (northern Chile): Quaternary Research, v. 52, p. 143–153, doi: 10.1006/qres.1999.2060.

Gile, L.H., Peterson, F.F., and Grossman, R.B., 1966, Morphological and genetic sequences of carbonate accumulations in desert soils: Soil Science, v. 101, p. 347–360.

Grove, M.J., Baker, P.A., Cross, S.L., Rigsby, C.A., and Seltzer, G.O., 2003, Application of strontium isotopes to understanding the hydrology and paleohydrology of the Altiplano, Bolivia—Peru: Palaeogeography, Palaeoclimatology, Palaeoecology, v. 194, p. 281–297, doi: 10.1016/S0031-0182(03)00282-7.

Haberle, S.G., and Maslin, M.A., 1999, Late Quaternary vegetation and climate change in the Amazon Basin based on a 50,000 year pollen record from the Amazon Fan, ODP Site 932: Quaternary Research, v. 51, p. 27–38, doi: 10.1006/qres.1998.2020.

Harris, S.E., and Mix, A.C., 1999, Pleistocene precipitation balance in the Amazon Basin recorded in deep sea sediments: Quaternary Research, v. 51, p. 14–26, doi: 10.1006/qres.1998.2008.

Haug, G.H., Hughen, K.A., Sigman, D.M., Peterson, L.C., and Rohl, U., 2001, Southward migration of the Intertropical Convergence Zone through the Holocene: Science, v. 293, p. 1304–1308, doi: 10.1126/science.1059725.

Hillaire-Marcel, C., Odette, C., and Casanova, J., 1986, ^{14}C and Th/U dating of Pleistocene and Holocene stromatolites from East African paleolakes: Quaternary Research, v. 25, p. 312–329, doi: 10.1016/0033-5894(86)90004-9.

Kastner, T.P., and Goñi, M.A., 2003, Constancy in the vegetation of the Amazon Basin during the late Pleistocene: Evidence from the organic matter composition of Amazon deep sea fan sediments: Geology, v. 31, p. 291–294, doi: 10.1130/0091-7613(2003)031<0291:CITVOT>2.0.CO;2.

Lao, Y., and Benson, L., 1988, Uranium-series age estimates and paleoclimate significance of Pleistocene tufas from the Lahontan Basin: California and Nevada: Quaternary Research, v. 30, p. 165–176.

Laskar, J., 1990, The chaotic motion of the solar system: A numerical estimate of the chaotic zones: Icarus, v. 88, p. 266–291, doi: 10.1016/0019-1035(90)90084-M.

Lavenu, A., Fornari, M., and Sebrier, M., 1984, Existence de deux nouveaux épisodes lacustres Quaternaires de l Altiplano Péruvo-Bolivien: Cahiers de l'ORSTOM, Série Geologie, v. 14, p. 103–114.

Lenters, J.D., and Cook, K.H., 1997, On the origin of the Bolivian high and related circulation features of the South American climate: Journal of Atmospheric Sciences, v. 54, p. 656–677, doi: 10.1175/1520-0469(1997)054<0656:OTOOTB>2.0.CO;2.

Malek, E.A., 1985, Snail host of Schitosomiasis and other snail-transmitted diseases in tropical America: A manual: Pan American Health Scientific Publication 478, p. 75–77.

Minchin, J., 1882, Notes of a journey through part of the Andean table-land of Bolivia in 1882: Proceedings of the Royal [London] Geographic Society, v. 4, p. 671–676.

Morrison, R., 1978, Quaternary soil stratigraphy—Concepts, methods, problems, *in* Mahaney, W.C., ed., Quaternary soils: Norwich, Geoabstracts, p. 77–108.

Mourguiart, P., and Ledru, M.P., 2003, Last Glacial Maximum in an Andean cloud forest environment (eastern Cordillera, Bolivia): Geology, v. 31, p. 195–198, doi: 10.1130/0091-7613(2003)031<0195:LGMIAA>2.0.CO;2.

North American Commission on Stratigraphic Nomenclature, 1983, North American stratigraphic code: American Association of Petroleum Geologists Bulletin, v. 67, p. 841–875.

Oviatt, C.G., McCoy, W.D., and Nash, W.P., 1994, Sequence stratigraphy of lacustrine deposits: A Quaternary example from the Bonneville Basin, Utah: Geological Society of America Bulletin, v. 106, p. 133–144, doi: 10.1130/0016-7606(1994)106<0133:SSOLDA>2.3.CO;2.

Palmer, M.R., and Pearson, P.N., 2003, A 23,000-year record of surface water pH and pCO_2 in the western equatorial Pacific Ocean: Science, v. 300, p. 480–482, doi: 10.1126/science.1080796.

Petit, J.R., Jouzel, J., Raynaud, D., Barkov, N.I., Barnola, J.-M., Basile, I., Chappellaz, J., Davis, M., Delaygue, G., Delmotte, M., Kotlyakov, V.M., Legrand, M., Lipenkov, V.Y., Lorius, C., Pépin, L., Ritz, C., Saltzman, E., and Stievenard, M., 1999, Climate and atmospheric history of the past 420,000 years from the Vostok ice core, Antarctica: Nature, v. 399, p. 429–436, doi: 10.1038/20859.

Pigati, J.S., Quade, J., Shanahan, T.M., and Haynes, V.C., Jr., 2004, Radiocarbon dating of minute gastropods and new constraints on the timing of late Quaternary spring-discharge deposits in southern Arizona, USA: Palaeogeography, Palaeoclimatology, Palaeoecology, v. 204, p. 33–45, doi: 10.1016/S0031-0182(03)00710-7.

Placzek, C., Patchett, P.J., Quade, J., and Wagner, J.D.M., 2006, Strategies for successful U-Th dating of paleolake carbonates: An example from the Bolivian Altiplano: Geochemistry, Geophysics, Geosystems, v. 7 (in press).

Risacher, F., and Fritz, B., 2000, Bromine geochemistry of Salar de Uyuni and deeper salt crusts, central Altiplano, Bolivia: Chemical Geology, v. 167, p. 373–392, doi: 10.10161/S0009-2541(99)00251-X.

Roche, M.A., Bourges, J., Cortes, J., and Mattos, R., 1992, Climatology and hydrology of the Lake Titicaca Basin, *in* Dejoux, C., and Iltis, A., eds., Lake Titicaca: A synthesis of limnological knowledge: Dordrecht, Kluwer Academic Publishers, p. 63–88.

Rondeau, B., 1990, Géochemie istopique et géochronologie des stromatolite lacustres quaternaries de l'Altiplano Bolivien [Ph.D. thesis]: Montreal, Canada, Université du Québec à Montréal, 100 p.

Rouchy, J.M., Servant, M., Fournier, M., and Causse, C., 1996, Extensive carbonate algal bioherms in Upper Pleistocene saline lakes of the central Altiplano of Bolivia: Sedimentology, v. 43, p. 973–993.

Rowe, H.D., Dunbar, R.B., Mucciarone, D.A., Seltzer, G.O., Baker, P.A., and Fritz, S., 2002, Insolation, moisture balance and climate change on the South American Altiplano since the Last Glacial Maximum: Climatic Change, v. 52, p. 175–199.

Samos, G., 1949, Some observations on exchange of CO_2 between $BaCO_3$ and CO_2 gas: Science, v. 110, p. 663–665.

Seltzer, G., Rodbell, D., and Burns, S., 2000, Isotopic evidence for late Quaternary climatic change in tropical South America: Geology, v. 28, p. 35–38, doi: 10.1130/0091-7613(2000)028<0035:IEFLQC>2.3.CO;2.

Servant, M., and Fontes, J.C., 1978, Les lacs quaternaries des hautes plateaux des Andes Boliviennes: Cahiers de l'ORSTOM: Série Géologie, v. 10, p. 79–97.

Servant, M., and Servant-Vildary, S., 2003, Holocene precipitation and atmospheric changes inferred from river paleowetlands in the Bolivian Andes: Palaeogeography, Palaeoclimatology, Palaeoecology, v. 194, p. 187–206, doi: 10.1016/S0031-0182(03)00277-3.

Steinmann, G., Hok, H., and Bistram, A., 1904, Zur geoligie des suedestland Bolivien: Zentralblatt für Mineralogie, v. 5, p. 1–4.

Stuiver, M., and Reimer, P.J., 1993, Extended ^{14}C data base and revised CALIB 3.0 ^{14}C age calibration program: Radiocarbon, v. 35, p. 215–230.

Stute, M., Forster, M., Frischkorn, H., Serejo, A., Clark, J.F., Schlosser, P., Broecker, W.S., and Bonani, G., 1995, Cooling of tropical Brazil (5°S) during the Last Glacial Maximum: Science, v. 269, p. 379.

Sylvestre, F., 2002, A high-resolution diatom reconstruction between 21,000 and 17,000 ^{14}C yr BP from the southern Bolivian Altiplano (18–23°S): Journal of Paleolimnology, v. 27, p. 45–57, doi: 10.1023/A:1013542907394.

Sylvestre, F., Servant, M., Servant-Vildary, S., Causse, C., Fournier, M., and Ybert, J.P., 1999, Lake-level chronology on the southern Bolivian Altiplano (18–23°S) during late-glacial time and the early Holocene: Quaternary Research, v. 51, p. 54–66, doi: 10.1006/qres.1998.2017.

Thompson, L.G., Mosley-Thompson, E., Davis, M.E., Lin, P.-N., Henderson, K.A., Cole-Dai, J., Bolzan, J.F., and Liu, K.-B., 1995, Late glacial stage and Holocene tropical ice core records from Huascarán, Peru: Science, v. 269, p. 46–50.

Thompson, L.G., Davis, M.E., Mosley-Thompson, E., Sowers, T.A., Henderson, K.A., Zagorodnov, V.S., Lin, P.-N., Mikhalenko, V.N., Campen, R.K., Bolzan, J.F., Cole-Dai, J., and Francou, B., 1998, A 25,000-year tropical climate history from Bolivian ice cores: Science, v. 282, p. 1858–1864, doi: 10.1126/science.282.5395.1858.

Vuille, M., 1999, Atmospheric circulation over the Bolivian Altiplano during dry and wet periods and extreme phases of the Southern Oscillation: International Journal of Climatology, v. 19, p. 1579–1600, doi: 10.1002/(SICI)1097-0088(19991130)19:14<1579::AID-JOC441>3.0.CO;2-N.

Vuille, M., and Keimig, F., 2004, Interannual variability of summertime convective cloudiness and precipitation in the central Andes derived from ISCCP-B3 data: Journal of Climate, v. 17, p. 3334–3348, doi: 10.1175/1520-0442(2004)017<3334:IVOSCC>2.0.CO;2.

Vuille, M., Hardy, D.R., Braun, C., Keimig, F., and Bradley, R.S., 1998, Atmospheric circulation anomalies associated with 1996/1997 summer precipitation events on Sajama ice cap, Bolivia: Journal of Geophysical Research, v. 103, p. 11,191–11,204, doi: 10.1029/98JD00681.

Wang, X., Auler, A.S., Edwards, R.L., Cheng, H., Cristalli, P.S., Smart, P.L., Richards, D.A., and Shen, C.-C., 2004, Wet periods in northeastern Brazil over the past 210 kyr linked to distant climate anomalies: Nature, v. 432, p. 740–743, doi: 10.1038/nature03067.

Wirrmann, D., and Mourguiart, P., 1995, Late Quaternary spatio-temporal limnological variations in the Altiplano of Bolivia and Peru: Quaternary Research, v. 43, p. 344–354, doi: 10.1006/qres.1995.1040.

Zhou, J., and Lau, K.-M., 1998, Does a monsoon climate exist over South America?: Journal of Climate, v. 11, p. 1020–1040, doi: 10.1175/1520-0442(1998)011<1020:DAMCEO>2.0.CO;2.

Manuscript Received by the Society 1 December 2004
Revised Manuscript Received 24 May 2005
Manuscript Accepted 1 October 2005

Printed in the USA